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A geologic map, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, may be obtained. Price 25 cents. Remittance should accompany order.

BIBLIOGRAPHY AND INDEX OF LITERATURE ON URANIUM AND THORIUM AND RADIOACTIVE OCCURRENCES IN U. S. PART 4—ARKANSAS, IOWA, KANSAS, LOUISIANA, MINNESOTA, MISSOURI, NEBRASKA, NORTH DAKOTA, OKLAHOMA, SOUTH DAKOTA, TEXAS. By Margaret Cooper

DRUMLINS WITH JOINTED BOULDERS NEAR DOLLARD, SASKATCHEWAN, CANADA. By W. O. Kupsch

EXPERIMENTAL ANALYSIS OF FRACTURE PATTERNS. By Ernst Cloos

GEOMORPHOLOGY OF SOUTH-CENTRAL WASHINGTON, ILLUSTRATED BY THE YAKIMA EAST QUADRANGLE. By Aaron C. Waters

A geologic map, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, may be obtained. Price 25 cents. Remittance should accompany order.

GEOLOGY OF THE LIMA REGION, SOUTHWESTERN MONTANA AND ADJACENT IDAHO. By Robert Scholten, K. A. Keenmon, and W. O. Kupsch

A geologic map, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, may be obtained. Price 25 cents. Remittance should accompany order.

DISTRIBUTION OF IGNEOUS ROCKS IN SPACE AND TIME. By G. W. Tyrrell

PROVENIENCE OF PYROCLASTIC ROCKS. By Clarence S. Ross

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SOURCE OF SHARON CONGLOMERATE OF NORTHEASTERN OHIO

BY J. OSBORN FULLER[1]

ABSTRACT

Sharon conglomerate crops out in three separate areas in Ohio and in all of these the formation has similar characteristics and is the basal Pennsylvanian bed resting unconformably on Mississippian sandstone or shale. This paper is concerned chiefly with the larger area of outcrop in northeastern Ohio which lies west of the type area at Sharon, Pennsylvania. The formation is dominantly a pure (96+% SiO_2), medium-grained, quartz sandstone (orthoquartzite) with a few pebble layers or scattered pebbles. Laterally this rock grades into a series of narrow, conglomerate belts 1000± feet wide and 10–50 feet thick. Trend of these belts, and, in the whole formation, dip of cross-bedding, dip of the initial slope, direction of overturning of tops of cross-beds, imbricate structure, and north-south changes in sedimentary features, mechanical analyses, and chemical analyses all indicate a northern source for the clastic material of the Sharon conglomerate and also suggest a deltaic type of depositional environment. In older literature the pebbles of the Sharon are described as "vein quartz" suggesting an igneous rock source. Fossils as pebbles, and pebbles of sandstone, and conglomerate indicate rocks that were in part sedimentary. Almost no minerals are present except quartz and most of the quartz grains show secondary growth, indicating that the individual grains have been through more than one cycle of sedimentation. Although igneous rocks were doubtlessly the original source of many of the multicycle particles, the lack of pebbles of igneous rock and grains of igneous rock accessory minerals strongly suggests that the sediments of the Sharon are several cycles removed from the original source rocks. This concept is confirmed by examination of thin sections. Most pebbles show ghost granular structure and the inclusions in both pebbles and sand grains are characteristic of quartz derived from a metamorphic source.

The fossils, most of which are tabulate corals, prove that at least part of the source was Middle Devonian (Onondaga-Hamilton). The writer concludes that the Sharon sediments originated in a Canadian highland composed of a mixture of sedimentary and metamorphic rocks some of which were of Middle Devonian age, and that these sediments were transported by streams of considerable competency and were ultimately deposited as a delta in a shallow basin developed on Mississippian rocks.

The characteristics of the Sharon in southern Ohio indicate a similar origin but the clastic components were apparently derived from the southeast as shown by direction of dip of cross-beds and trend of conglomerate belts.

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[1] Presented with the encouragement of the Geological Survey of Ohio.

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Introduction

The basal Pennsylvanian Sharon conglomerate of Ohio lies disconformably on the even-bedded marine Mississippian shale and sandstone, in a relationship similar to its correlatives, the Olean of New York and the Mansfield of Indiana. In Ohio the Sharon conglomerate crops out in three separate areas (Fig. 1). Area A covers about 150 square miles in Ohio and is a continuation from the type area at Sharon, Mercer County, Pennsylvania. Area B spreads over about 2400 square miles. It is separated from Area A by the 12-mile-broad Grand River Valley which has been cut in Mississippian rocks. Area C covers about 450 square miles in south-central Ohio. It is separated from Areas A and B by an area about 100 miles along the strike in which the Sharon is lacking and the basal Pennsylvanian is a younger formation than the Sharon. This paper is chiefly concerned with Area B in which mapping has been most detailed.

Acknowledgments

Work on the Sharon conglomerate was sponsored by the Geological Survey of Ohio which is interested in encouraging further exploitation of this pure silica rock already important for foundry sand, ferrosilicon and silica brick. The writer has worked on the Sharon every summer since 1941 except for an interruption during the war. He acknowledges his indebtedness to the three state geologists of Ohio under whom this work has been done: Dr. Wilber Stout, Dr. George W. White and Mr. John H. Melvin. Dr. Myron T. Sturgeon of Ohio University worked on the problem with the writer and independently during the war; he deserves credit for some of the early discoveries concerning the nature of the Sharon. The writer is also indebted to his field assistants: Dr. William M. Merrill, Dr. Charles N. Bowen, Mr. Robert E. Lehner and Mr. John J. Pedry. Drs. George W. White and J. A. Aurelé LaRocque criticised the manuscript for the writer.

Character of Sharon in Area B

Orthoquartzite

Sharon conglomerate of Area B crops out extensively in Geauga, Portage, Summit and Medina counties, and sparsely in the surrounding counties, Trumbull, Lake, Cuyahoga and Wayne (Fig. 1). In this area of about 2400 square miles the Sharon is dominantly an orthoquartzite. Average grain size is 0.25–0.5 mm. A few scattered pebbles and pebble lenses are common and grit layers are rare. Sand grains characteristically flash light from many crystal faces of secondary quartz. Normally the Sharon is a clean, white, friable orthoquartzite with a silicon dioxide content of over 96 per cent; near the surface and along joint planes it is limonite-stained and more solidly cemented. Next to limonite the chief impurities are clay and feldspar. Thickness throughout this vast sheet of orthoquartzite varies from 0 to 170 feet. Cross-bedding is abundant and ripple marks are present locally.

Conglomerate Belts

Long narrow belts of conglomerate with a southerly trend lie in the orthoquartzite. They are difficult to trace because of cover and because their lateral boundaries are indistinct. The most distinctive belt extends south of Thompson, Ohio for seven miles (Fig. 2-1). Conglomerate appears 17½ miles south at Nelson Ledges (Fig. 2-2) and extends slightly east of south for 4 miles. Position and alignment suggest a possible connection at one time between these two belts but this has not been definitely established. Two other belts are

fairly well defined, the first extending northeast for 4 miles from the Best quarry located south of Chardon (Fig. 2-3) and the second extending for 3 miles southwest of the Copley quarry (Fig. 2-4). Other short linear strips of conglomerate have been found (Fig. 2). The pattern and alignment of all the conglomerate belts suggest distributary channels on a deltaic or alluvial plain.

Along a conglomerate belt the material is well sorted and uniform but across a conglomerate belt in 500–1000 feet the rock grades from a conglomerate into an orthoquartzite with scattered pebbles and then into an almost pebble-free orthoquartzite. The gradational area frequently has lenses of orthoquartzite in the conglomerate. The long axis of some of these lenses is parallel to the bedding (Pl. 1, fig. 1) and in others cuts across the bedding. These lenses are believed to be remnants of sandbars and similar features, and indicate rapidly changing conditions of deposition and erosion. They are more abundant in the southern part of the area than in the northern part.

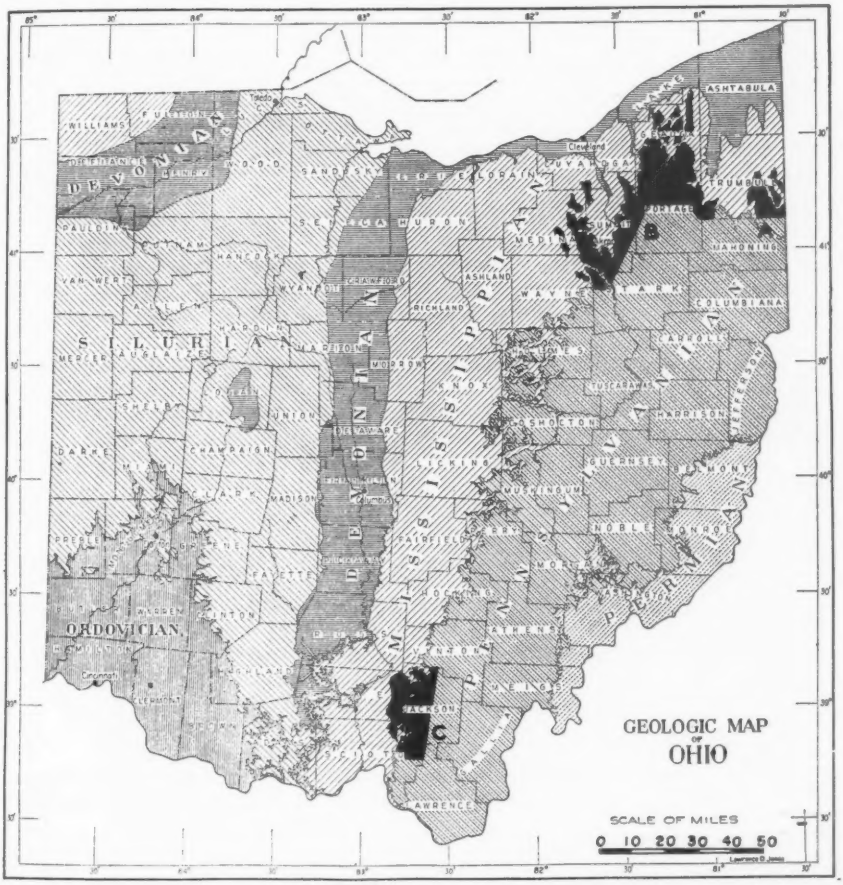

FIGURE 1.—LOCATION OF THE THREE PRINCIPAL AREAS OF OUTCROP OF SHARON CONGLOMERATE IN OHIO

Area A contiguous with the type area at Sharon, Pennsylvania; Area B the area of this report; Area C Sharon of southern Ohio

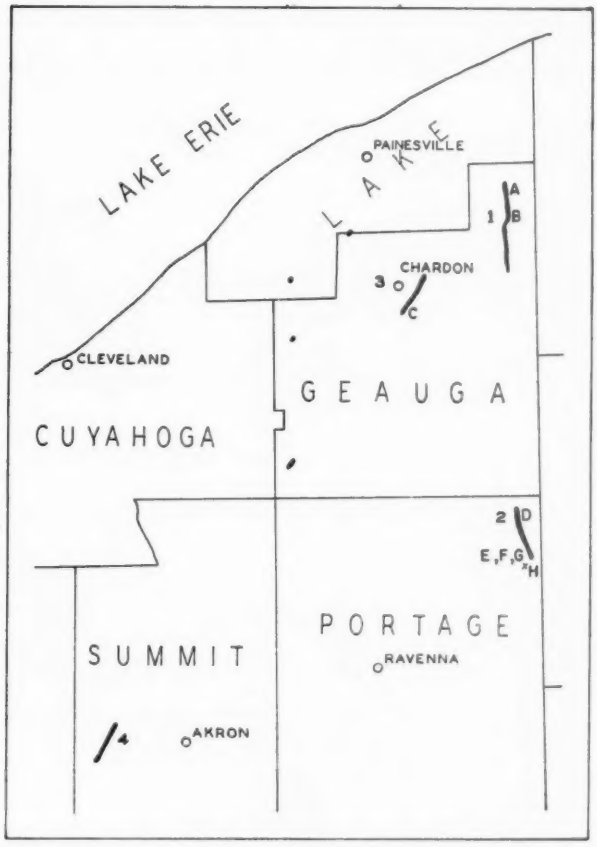

FIGURE 2.—DISTRIBUTION OF KNOWN CONGLOMERATE BELTS IN THE SHARON

Heavy solid lines indicate position of thick conglomerate sections. Distinctive belts are: Thompson (1), Nelson (2), Chardon (3), and Copley (4). Letters indicate approximate locations of specimens with chemcal analyses shown in Table 1.

In an alluvial plain or delta deposit, deposition is more variable near the outer limit than near the source. If the Sharon sheet represents either type of deposit, the more abundant orthoquartzite lenses in the south would suggest a northern source for the Sharon.

PLATE 1.—SEDIMENTARY FEATURES OF THE SHARON CONGLOMERATE

FIGURE 1.—ORTHOQUARTZITE LENS IN CONGLOMERATE PHASE OF SHARON CONGLOMERATE
Lens lies at an angle to the principal bedding

FIGURE 2.—CHANNEL CUT IN ORTHOQUARTZITE PHASE OF THE SHARON AND FILLED WITH CONGLOMERATE PHASE

FIGURE 3.—CLAY MASSES IN THE CONGLOMERATE PHASE OF THE SHARON
Large clay mass above the hammer shows no contortion and little rounding of the end; clay mass at hammer point was contorted and rounded during movement.

FIGURE 4.—SHARON CONGLOMERATE SHOWING LAG GRAVELS
Bed below hammer with faint cross-bedding is typical of orthoquartzite with scattered pebbles; below this bed and above the hammer are beds of lag gravels.

FIGURE 1

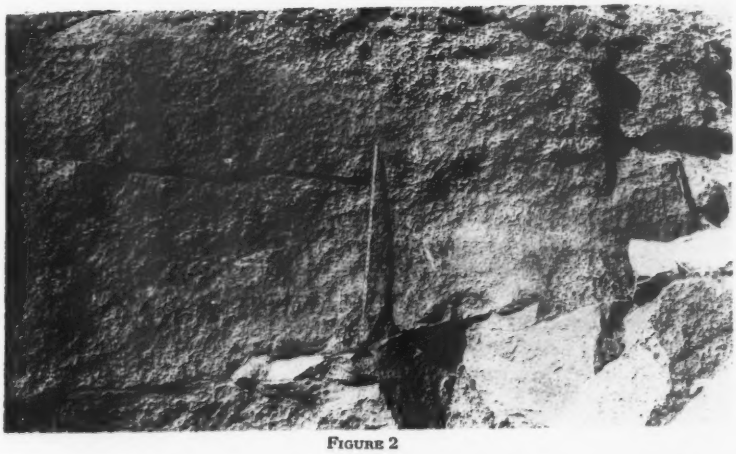

FIGURE 2

FIGURE 3

FIGURE 4

SEDIMENTARY FEATURES OF THE SHARON CONGLOMERATE

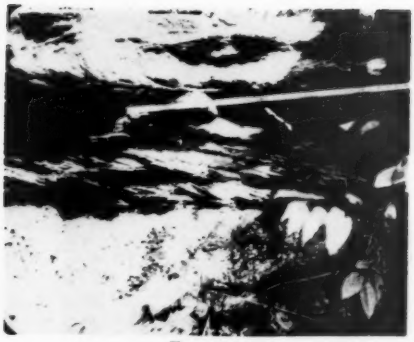

FIGURE 1

FIGURE 2

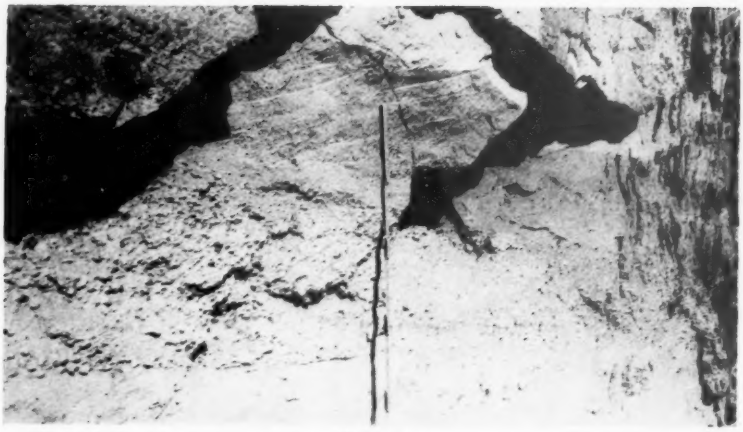

FIGURE 3

FIGURE 4

CHARACTERISTICS OF THE SHARON CONGLOMERATE

Some orthoquartzite-conglomerate contacts are sharp, where the conglomerate shows a definite channel relationship to orthoquartzite or orthoquartzite with scattered pebbles (Pl. 1, fig. 2). Other contacts are gradational, with the conglomerate gradually changing into a pebbly orthoquartzite and sometimes into a pebble-free orthoquartzite. These lateral variations suggest scour channels, the definite contacts being on the undercut side of the stream and the gradational contacts on the slipoff slopes.

Shale lenses and clay masses are associated with the conglomerate belts. Shale and clay are blue gray in color and weather light gray; both are slightly sandy. Lenses of shale a few inches to several feet thick are visible for several hundred feet in some outcrops. Clay masses are either angular pieces, which show little evidence of movement except their random orientation in the deposit, or rounded masses with contorted bedding, obviously molded during transportation (Pl. 1, fig. 3). These masses range from pieces a fraction of an inch in diameter to lenticular bodies 6 feet long and 1 foot thick.

The lenses appear to be quiet water deposits formed away from the main currents. When the main current would shift back into the area, it would either cover the fine clastics with sand and pebbles forming a shale lens or sweep up blocks of clay some of which were moved intact and others contorted and rounded during transportation.

Conglomerate Beds

Several horizontal beds of well-sorted conglomerate are usually seen in large exposures of orthoquartzite or conglomerate. When present in orthoquartzite these beds are very distinctive and can be followed across the whole outcrop although they may be only one pebble thick or several feet thick. Even in an outcrop which is dominantly conglomerate, these beds are distinctive because their pebbles are so well sorted while those of adjacent beds are more poorly sorted (Pl. 1, fig. 4). Flattened pebbles in these beds are often found imbricated with an alignment indicating a northern source.

Gradient apparently was such that the characteristic deposit was a mixture ranging from fine sand to coarse gravel. Probably the sand moved by saltation and in suspension while the pebbles rolled along the bottom. If the current increased and caused erosion of this newly deposited sand and gravel mixture, the sand would be swept free and the pebbles concentrated slightly lower in a layer. These well-sorted, horizontal, conglomerate beds, therefore, are called lag gravels.

Cross-Stratification

Both orthoquartzite and conglomerate show abundant cross-stratification. In describing this feature the writer has followed the purely descriptive classification of McKee and Weir (1953, p. 381–390), even though he does not agree completely with them. The strata of the cross-beds dip south and one planar cross-set is piled on top of another with occasional intervening strata with horizontal bedding.

Each stratum is normally between one half and two inches thick though thicker strata have been observed. According to McKee (1953, p. 383) the rock would be classified as very thinly cross-bedded. In the grits and conglomerates the cross-bedding stands out be-

PLATE 2.—CHARACTERISTICS OF THE SHARON CONGLOMERATE

FIGURE 1.—CROSS-BEDS WITH OVERTURNED TOPS

The upper bed shows cross-bedding whose top has been overturned due to creep on the original depositional slope. Right is south.

FIGURE 2.—GRADED CROSS-BEDDING

Phalanx quarry of Industrial Silica Corp.

FIGURE 3.—INITIAL DEPOSITIONAL SLOPE TO THE SOUTH

Bedding planes separating cross-bedded conglomerate layers show initial depositional slope to south (left) which has been slightly accentuated by later orogenic movement.

FIGURE 4.—COMMON OCCURRENCE OF CONGLOMERATE NEAR BASE OF SHARON WITH THE SANDSTONE PHASE OVER IT

Mississippian-Pennsylvanian contact just below the quarry base

cause the textures of the successive strata differ. Texture variation probably also accounts for the cross-bedding in the medium-grained orthoquartzites although this cannot be determined by the naked eye. The cross-beds may be a set of parallel foreset beds with no top or bottomsets but more characteristically they are a set of foresets grading imperceptibly into bottomsets, the whole with a slight upward concavity. No topset beds have been observed but some of the horizontal layers probably represent reworked ones; if the reworking continues long enough a well-sorted pebble layer results which has been described as a lag gravel.

In weathered sections the cross-bedding planes may serve as bounding surfaces and flaggy blocks may develop. Where a stream cuts across the dip of these blocks a confusing series of false bedding planes results. The cross-strata are of medium scale with an average length of several feet, but some are large scale—20–45 feet long.

A set of cross-strata is tabular and may be followed for several hundred feet along a quarry face or outcrop with little change in thickness. Eventually it thins and disappears owing to erosion and is replaced by another set of cross-strata or by horizontal layers. The thickness of the set is normally 1 foot to 1½ feet and the range is 6 inches to 6 feet.

Locally, erosion has cut troughs through the planar cross-bedding and the deposits in these troughs are themselves cross-bedded. Except for the curved erosion surface at the base, this cross-bedding is similar in every way to the planar cross-bedding, and would be classified by McKee and Weir as trough cross-bedding. It is more prevalent in the southwestern part of the area, indicating that conditions changed more rapidly in the south, and further supporting the concept of a northern source.

Another feature of some cross-bedded areas is the overturning of the top of the cross-beds, which gives the cross-beds the appearance of being folded (Pl. 2, fig. 1). This apparent folding is not an orogenic feature because it lies between undisturbed horizontal layers. It is obviously caused by creep of the cross-bedded sand down the depositional slope before consolidation and before deposition of the overlying beds. All observed overturnings are southerly in direction.

Several hundred measurements were taken of the strike and dip of the cross-beds. In the Sharon the dip directions range from east through south to west. The dip varies from 7° to 26° S. with the greatest number 20° ± a few degrees. In a single outcrop several dip directions for the cross-beds might be observed, but they clustered around a general trend—for example, all dip directions might be in the southeast quadrant. Lack of northeast and northwest dipping cross-beds indicates that the flowing water from which these cross-beds were deposited must have had a northern source.

In one locality in the southeastern part of the outcrop area (Industrial Silica Corp. Phalanx quarry) graded cross-bedding is observed in some strata (Pl. 2, fig. 2). This phenomenon is due to deposition of cross-beds in quiet water. More rapid settling of the coarser material caused the development of graded bedding in individual cross-strata.

Initial Slopes

In most large outcrops of the Sharon the tops of the cross-bedded layers and the associated relatively horizontal beds dip about 3° S. or about 275 feet per mile (Pl. 2, fig. 3). On the basis of a study of the Berea of the eastern part of this region, De Witt (1946) drew structure contours on top of the Berea which indicate an average southerly dip of 11 feet per mile. The difference between the amount of regional dip and the dip observed in the beds of Sharon indicates that the Sharon built forward with a marked initial slope. The initial slope varies in direction from southeast to southwest normally, but in one locality, the east side of the number 1 quarry of the Industrial Silica Corp. at Phalanx, there is a dip 4° E. This seems to be a depositional slope where the deposits were building up against a Mississippian shale hill.

Composition

The high silica content of the Sharon orthoquartzite and conglomerate has been mentioned. Bowen (1953, p. 41–43) published the

first modern analyses of the Sharon. He cut clean channel samples from the whole exposed cliff face of the quarries in the northern part of Area B. The samples represent the total thickness of the formation in these localities lacking only a few feet which have been removed by erosion or left in the quarry bottom for a floor. Because the depth of the sample taken was about one inch in sandy material and equal to the diameter of the largest pebble in conglomeratic material the total sample ranged in weight between 10 and 50 pounds. In a formation which varies rapidly in texture both horizontally and vertically these clean samples cutting across practically the whole formation give a better concept of the composition than was available from the grab samples used in older analyses.

Bowen (1953, p. 41–43) included in his Table II analyses of two samples which were shown by Pedry (1951, thesis Ohio State University, p. 57–58) to be Massillon instead of Sharon. Elimination of these two samples leaves eight certain Sharon analyses, six of which are conglomerates and two orthoquartzites. (See Table 1.) The last two columns of Table 1 give the averages of the conglomerates and orthoquartzites respectively. Both averages indicate that Sharon SiO_2 content is in the high nineties and that the only impurities of importance are Al_2O_3 and Fe_2O_3. The next most abundant impurity is TiO_2 which averages 0.07 in the eight samples. This average is not very representative because only one sample has over 0.07 TiO_2. All other impurities are insignificant.

Comparison of the average analyses of conglomerates and orthoquartzites confirms two field observations regarding purity: that the conglomerates are slightly purer than the orthoquartzites and that this difference is mainly due to a higher clay content in the fines.

Bowen's analyses of the plus-four-mesh compared with the minus-four-mesh of the same sample confirms the idea that impurities

TABLE 1.—ANALYSES OF SHARON SAMPLES ARRANGED FROM NORTH TO SOUTH TO SHOW SYSTEMATIC CHANGE

	Conglomerates						Orthoquartzites		Average of Conglomerates	Average of Orthoquartzites
	A	B	C	D	E	F	G	H		
SiO_2	98.95	98.80	98.13	98.82	97.87	97.81	97.91	96.62	98.38	97.27
R_2O_3* group	0.74	0.92	1.15	0.72	1.45	1.44	1.59	2.23	1.07	1.91
Al_2O_3	0.52	0.63	0.85	0.38	0.88	0.92†	1.24	1.72	0.71	1.48
Fe_2O_3	0.17	0.24	0.24	0.31	0.50	0.45	0.29	0.38	0.32	0.34
TiO_2	0.04	0.03	0.05	0.02	0.06	0.05	0.05	0.12	0.04	0.09
MnO	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01
CaO	0.00	0.00	0.18	0.00	0.06	0.11	0.04	0.00	0.06	0.02
MgO	0.02	0.02	0.01	0.01	0.03	0.01	0.03	0.08	0.02	0.06
K_2O	0.04	0.05	0.05	0.04	0.08	0.03	0.09	0.24	0.05	0.17
Na_2O	0.01	Tr	0.00	0.00	0.01	0.00	0.01	0.00	0.01	0.01
SO_3	0.01	0.01	0.06	0.00	0.00	0.05	0.00	0.00	0.02	0.00

* R_2O_3 group contains Al_2O_3, Fe_2O_3, FeO, FeS_2, TiO_2, MnO, V_2O_5, P_2O_5 and ZrO_2.

† By mistake given as 0.97 in Table II of Bowen (personal communication).

A—R. W. Sidley Inc. Thompson Quarry.

B—R. W. Sidley Inc. South Thompson Quarry.

C—Walter C. Best Inc.

D—Harbison-Walker Refractories Co., Nelson Ledges.

E—Industrial Silica Corp. Phalanx Plant.

F—Industrial Silica Corp. Phalanx Plant.

G—Industrial Silica Corp. Phalanx Plant.

H—Industrial Silica Corp. Sandy Flats Quarry.

are more abundant in the finer material. He states,

"The plus-four-mesh grade varied between 98.92 per cent and 99.59 per cent silica, the minus-four-mesh grade from 96.62 per cent to 98.69 per cent silica. The chief contaminating ingredient was alumina which varied between 0.06 and 0.34 per cent in the coarse fraction and from 0.66 to 1.72 in the fine." (Bowen, 1953, p. 39.)

In Table 1 the eight Sharon analyses are arranged in order from north to south. Position of the localities is shown on Figure 1. Samples E, F and G are from different parts of the Industrial Silica Corporation's quarry at Phalanx and were taken roughly east-west across a conglomerate belt. Sample E is conglomerate with very little sand matrix; Sample F is conglomerate with a fair percentage of sand matrix, and Sample C is an orthoquartzite with a few scattered pebbles. Variations in chemical composition between these three specimens are probably due to increased fineness of particles and consequent increase in impurity. Although E is actually north of F and F north of G they should be treated as one locality when considering the north-south variation in composition. The significant horizontal lines in the table are the first four. The SiO_2 content decreases progressively from north to south while the impurities represented by the other three columns gradually increase. The progression is not perfect but it seems significant. The number of samples considered is so small that this change in composition from north to south is not sufficient in itself to prove a northern source for the Sharon conglomerate but it does advance confirmation.

Bownocker (1921; 1926) published the only other analyses of the Sharon. Several samples he called Sharon are now known to be Massillon and the location of others is so indefinite that they may be either Massillon or Sharon. All his samples were chip samples and some were iron-stained. Yet six certain Sharon samples (1921, p. 256; 1926, p. 39–40) are from locations in the southwestern part of the area where Bowen did not work; their average composition is: SiO_2 98.17; Al_2O_3 0.43; Fe_3O_4 0.39; CaO 0.03; MgO 0.06; TiO_2 0.05; loss on ignition 0.32. This substantiates the field observation that the Sharon throughout Area B is a pure orthoquartzite or conglomerate.

Mechanical Analyses

Bowen mechanically analyzed his samples from the northeastern part of Area B. He reached several conclusions which are in accord with the ideas previously presented by the writer (Fuller 1947; 1950a; 1950b) and expanded in this paper.

Plotting by Bowen (1952, dissertation, Ohio State University, p. 60–61) of median grain diameter against logarithmic sorting coefficient indicates better sorting in samples with smaller median diameter than in the coarser samples. This agrees with the conclusion reached by Inman (1949, p. 67) for sediments transported by currents of water and supports the writer's idea that the Sharon is a water-laid deposit. Best sorting lies between 0.15 and 0.30 mm., which is higher than the 0.125–0.250 mm. stated by Inman as characteristic of well-sorted sediments. Bowen feels that this slight upward shift in size is probably related to post-depositional enlargement of the grains by silica and that the Sharon is a very well-sorted deposit with the characteristics of such deposits.

Cumulative curve plots of the median grain size against the sorting coefficient in the mechanical analysis of successive beds at a particular locality along a conglomerate belt produce a bundle of curves. The medians of these bundles show a progressive spread southward. The close spacing of the medians at the north "suggest that the competency of the currents in this part of the region tended to remain rather constant over a considerable period of time. The progressive widening of the curve bundles toward the south . . . indicates a marked fluctuation in competency at these sites." (Bowen, 1952, dissertation, Ohio State University, p. 62). The main current would be expected to meander more in the lower part of a delta or alluvial fan than in the upper part and this would account for fluctuation in competency at any one locality. Similarly, competency would fluctuate if the main current were subject to the braking power of a body of quiet water.

Bowen (1953, p. 11) reports that

"the percentage of pebble (plus-four-mesh) in the conglomerate showed a progressive decrease southward varying from 35 to 54 per cent pebble in the north to between 25 and 32 per cent in the south."

"A corresponding increase in sorting is noted as these belts are traced southward. These variations are quite similar to those noted in modern streams and indicate a movement of material from north to south." (Bowen, 1953, p. 47.)

In contrast to this gradual increase in sorting power along the channel axis, study of curves of samples taken across the axis in the Phalanx quarries of the Industrial Silica Corporation indicates a rapid increase in sorting with an equally rapid decrease in grain size. Bowen (1952, dissertation, Ohio State University, p. 68) notes that

"Russell (1939) has pointed out that progressive sorting according to size would be rapid where currents were carrying debris through a body of quiet water and with a rapid decrease in velocity. Hence, this rapid later increase in sorting suggests that the material away from the pebble channels was deposited in quiet water."

Bowen's works on mechanical analyses of the Sharon in the northeastern part of the area supports the idea of a northern source for the water-laid deposits of the Sharon. It also suggests that the progressive rapid sorting going outward from a channel center in the Phalanx region indicates deposition in a body of quiet water. As this is the same quarry in which graded cross-bedding is found, it seems that in this part of the region the Sharon sediments were deposited in a body of quiet water. These facts suggest deltaic plain environment rather than an alluvial fan.

Pebbles

In the early literature not much attention was paid to the character and significance of the pebbles in the Sharon conglomerate. They were usually identified as quartz, with occasional jasper pebbles. J. S. Newberry (1873, p. 213–214) published the most complete description of the pebbles and their origin, and the description is such a revelation of the thinking of the time that it is worth quoting in part:

"The pebbles of the more pebbly portions of the Conglomerate are sometimes as large as one's fist, but more generally range from the size of a hickory nut to that of an egg. They are almost always composed of quartz, but in every locality where they are abundant, more or less of them may be found which are composed of quartzite or silicious slate, which show lines of stratification. . . . There can be no question, however, that these pebbles are portions of quartz veins, which have been brought hundreds of miles from some area where metamorphic, crystalline rocks have suffered erosion. In the process of transportation the attrition to which these fragments were subjected comminuted all but the most resistant, viz: the quartz. The banded, silicious slates which are represented in the pebbles that accompany those of pure quartz, as well as the internal structure of the quartz pebbles themselves, afford conclusive evidence that their origin is such as I have described. . . . I have for many years been inclined to refer to transportation and deposition of the immense beds of quartz pebbles which are found in the Conglomerate to the same cause which has transported the gravels of the Drift, and the similar deposits which are now accumulating on the sea bottom off the Antarctic Continent, and on the Banks of New Foundland—viz: to ice."

Newberry's idea of ice transportation and deposition was soon dismissed by geologists when they learned more about the character of drift. His concept that "these pebbles are portions of quartz veins" was readily accepted and retained. Overlooking his reference to pebbles of metamorphic origin, later geologists spoke of the pebbles of the Sharon as being "vein quartz" derived from erosion of an "igneous source rock". Stout (1916, p. 441), in discussing the Sharon conglomerate of southern Ohio, considered the Sharon of northeastern Ohio and southern Ohio to have the same northern source. He says the Sharon "is largely crystallized quartz, and thus of igneous origin derived from the granitic rocks". Stout realizes the difficulty of obtaining all this quartz from erosion of a crystalline series but still holds to that concept (1916, p. 441). His (1916, p. 443) solution to the difficult problem involved in producing this amount of pure quartz is summarized as follows:

"Considering all points, the history of the Sharon conglomerate seems to indicate that the decay of the crystalline rocks reached an advanced state on low lying uplands; that the products were slowly transported to the sea, coarse materials lagging far behind the finer products; that the fine materials were distributed to the deep and quiet waters, forming shale, while the coarse materials were left more localized along the margins; that a general elevation took place, converting much of the floor of the sea to land, and restricting the currents to narrow shallow passages; that the force of these currents was now sufficiently strong to transport the coarse quartz, and thus build up the conglomerate."

Although Stout's explanation is a possibility, the following facts argue against it being the correct theory of origin:

1. Lack of any crystalline remnants, rocks or minerals except quartz

2. Well-rounded quartz grains and pebbles
3. Lack of extensive shale deposits represent-the fines deposited in deep water

In searching for the solution a detailed study of the quartz pebbles was started.

Over 90 per cent of the pebbles are a clear to milky quartz called "vein quartz" in the literature. The remaining pebbles described in the literature are smoky quartz, jasper, "silicious slates" and quartzites. The writer has found pebbles of sandstone, conglomerate, leached rotten limestone which has been silicified and several different orthoquartzites ranging from black to white. Most instructive of all was the discovery of quartzite conglomerates with pebbles just like those of the Sharon and the discovery of fossils as pebbles.

Most of the quartz pebbles, like the sand grains, show secondary enlargement by silica. These enlarged pebbles have a distinctive luster which is due to reflection of light from the many tiny crystal faces developed on the original curved surface.

Bowen (1953, p. 21) made an odd pebble collection at each locality he visited and states that the odd pebble suites "were purely qualitative in nature; however, they did indicate a progressive decrease in the more fragile types, sandstone, grainy quartzite, etc., from north to south." More resistant odd types like chert, jasper and well-cemented quartzite also decreased to the south but to a less marked degree. The north-south decrease in odd pebbles indicates longer transportation for the southern material and supports the contention of a northern source for the Sharon.

The pebbles of conglomerate are interesting because the sand matrix is just like the sand of the Sharon and the pebbles are the same as the white or milky quartz pebbles of the conglomerate phase of the Sharon. Even in a thin section no distinction could be made. Here then is a sedimentary remnant of an older formation which undoubtedly supplied some of the sand and pebble for the Sharon. Some of these pebbles are cut by tiny quartz veins, indicating intrusion of quartz into the sediments. Such veins may be the source of some milky quartz.

Bowen (1953, p. 10) discovered a glacial erratic resembling a Sharon boulder. Close examination indicated that it is different from true Sharon erratics. It is a coarse quartzite conglomerate, the quartz grains of the matrix are larger than the average grain size of the Sharon conglomerate and they are cemented with silica. Although Sharon conglomerate is sometimes well cemented by limonite, no siliceous cement has been found. The pebbles of the quartzite are the same as the more durable ones found in the Sharon, and in size are similar to the Sharon conglomerate pebbles although they are less well rounded and sorted and there is no cross-bedding. This boulder represents a formation exposed in Pleistocene time, and it is not unreasonable to believe that during Pennsylvanian time this formation may have been eroded to supply some of the materials for the Sharon.

Fossils

Fossils have been found only in the thicker conglomerate zones, but they have been discovered in five localities, the farthest separated being 40 miles apart. This suggests that the fossils are fairly widespread, and, now that the technique of locating them has been learned, examination of all conglomerate channels should turn up a few fossils.

In 1947 the best specimens of the first discovery were turned over to Dr. John W. Wells for identification. He reported:

"About 20 specimens of fossils, occurring as fragments in the Sharon conglomerate . . . have been examined . . . all are in poor condition mostly as a result of leaching and some silicification. All are incomplete and more or less worn, most of them [are] rugose or tabulate corals, but one or two fragments evidently were small bits of fossiliferous limestone and show small bits of trilobites, brachiopods, and *Tentaculites*. Unfortunately the trilobites and brachiopods are mostly quite indeterminable, and the only satisfactory identifications are for the corals and *Tentaculites*. List of specimens: Corals—*Favosites sp.* cf. *F. hemispherica* (Troost), *Favosites* sp. cf. *F. placenta* Rominger, *Emmonsia emmonsi* (M. E. & H.), *Emmonsia* sp., *Prismatophyllum* sp., *Synaptophyllum* sp. cf. *S. simcoense* (Billings); Brachiopod—*Ambocoelia umbonata* (Conrad)?; Gastropod—*Tentaculites bellulus* Hall. In addition to the above there are two specimens representing two additional genera of solitary rugose corals. The above material is all characteristically Middle Devonian (Onondaga-Hamilton). Extensive outcrops of strata of this age certainly existed during early Pottsville time to the east, north, and west, of the site of Sharon deposition, and the remanié specimens could have come from any or all of these. The matrix and fossils themselves, however, are too

altered to suggest a narrowing down of these sources on a lithologic basis."

Vein Quartz

Odd pebbles such as sandstones, conglomerates and fossils which were derived from sedimentary material establish the Sharon as partially of sedimentary source. However, the great bulk of the pebbles are milky or vitreous quartz and have been called "vein quartz", suggesting an igneous source. It should be noted that no pebbles of igneous rocks, of schist, or gneiss have been found. "Silicious slate" was reported by Newberry alone but he was probably referring to the dark-banded quartzite pebbles. Furthermore, the purity and secondary enlargement of quartz grains and pebbles indicate a deposit of at least a second generation. It seems logical, therefore, to think that a large percentage of these so called "vein quartz" pebbles might also be derived from a sedimentary source or from a pure metaquartzite.

It has already been pointed out that the conglomerate formation which supplied the conglomerate pebbles must have supplied some sand and pebble as well, and that some of the clear or milky quartz pebbles may have come from the source formation of the quartzite conglomerate erratic discovered by Bowen. That a pure metaquartzite was also a partial source must be considered.

Bowen examined about 20 thin sections of these pebbles and reported no positive grains of quartz. He states: "The so-called white massive pebbles appear to be vein or igneous quartz. All show large crystal size with a well developed lattice orientation. Many of the large crystals are in turn divided into smaller, subsidiary blocks or a mosaic pattern which the author regards as being due to strain." (Bowen, 1952, p. 92). While retaining the idea that the original source was igneous, Bowen admits that the purity of the whole formation signifies more than one cycle of deposition.

However, even if several cycles of erosion are accepted to justify the purity, the fact of an igneous origin is still not substantiated since not a single igneous pebble, no typical vein quartz accessories, and no variety to the heavy minerals are to be found. Erosion of a pure metaquartzite would be a more logical source for the Sharon quartz pebbles. An orthoquartzite could be metamorphosed so that almost all evidence of the former sedimentary character would be destroyed.

C. A. Lamey (1935) studied the Palmer gneiss which was previously considered part of the Archean Basement Complex cut by Laurentian granite. He found that it really consisted of a variety of Huronian sediments including quartzites of the Mesnard formation. The sericitic and ferruginous varieties of the quartzites usually show their clastic origin.

Lamey states (1935, p. 1154):

"The vitreous varieties, however, do not always so clearly show their clastic nature. This is usually revealed by a careful examination under a hand lens, and also by a microscopic inspection. Without the use of the analyzer the outlines of rounded quartz may be seen, especially if they are fringed with a small amount of iron oxide, but under crossed nicols these outlines, although observable, are not very apparent, due to the deposition of siliceous cement with the same optical orientation as the original quartz grains and in some instances there is little indication of original grains."

Pebbles derived from erosion of the vitreous quartzite described by Lamey could easily be mistaken for vein quartz.

Another example of vitreous quartz which is not igneous is the Sturgeon quartzite of the Felch Mountain range. Smyth, in a monograph by Clements and Smyth (1899, p. 400), describes the Sturgeon quartzite as a light-gray rock which breaks with a coarsely granular or glassy fracture. Quartz is often the only visibly recognizable constituent. The formation is so homogeneous that it is difficult to determine structure alone but field relations establish it as a sedimentary rock, well exposed in the Felch Mountain range in two parallel belts. Recrystallization has almost obliterated the sedimentary features and "Faint color banding, itself of secondary development, but no doubt preserving a distinction in original composition, alone remains, and only here and there, as a guide to the former stratification."

Smyth (*in* Clements and Smyth 1899, p. 402) describes a gradational series in thin sections from those consisting of large irregular grains of quartz with interlocking boundaries and inclusions of accessory minerals, through optically strained and elongated quartz, through fractured quartz with intergranular movement

producing a fine-grained quartz mosaic between the parted surfaces, and ending with highly strained quartz in long narrow lenses separated by anastomosing zones of finely divided quartz. All these were observed in individual pebbles of the Sharon. Erosion of a pure zone of quartzite similar to the Sturgeon quartzite could produce pebbles similar to the so-called "vein quartz" of the Sharon.

Since pure orthoquartzite can be metamorphosed so much that definite grain boundaries may be destroyed, causing it to resemble vitreous or vein quartz, the igneous quartz must be distinguished from metamorphic quartz on other criteria than definite dust rims or secondary quartz rims. Re-examination of Bowen's thin sections confirmed the lack of dust rims and clear boundaries of quartz grains. The slides show three types of granular pattern. A few show in plain light incipient fractures from stress which has broken crystals in small equigranular, angular blocks. Under crossed nicols all slides also show a granular pattern which represents the crystallographic units. All crystals have strain extinction. Their boundaries are interlocking in varying degrees of complexity and in a few specimens the boundaries are a mylonite between larger crystals. The third granular pattern is the most difficult to see but it is the most significant. In plain light, inclusions are found in all crystals in varying quantity. Where concentrated they are aligned to surround oval areas. In the lines where inclusions are most abundant there is also a greater variety of inclusions, some of which are crystals. Most are irregularly shaped gas or liquid inclusions with an index of refraction below quartz. The long axis of many irregular inclusions is on a tangent to the oval area of quartz. On the other hand, this type of inclusion within the oval mass is unoriented. In polarized light a single crystal is seen to span several of these rounded areas. Some are round while others appear elongated and angular. In a single pebble they are quite uniform in size and shape.

These oval areas are believed to represent original sand grains and the aligned inclusions are believed to be equivalent to the dust rims commonly observed when a metaquartzite has been produced. Because of high purity of the original sand and extreme recrystallization, grain boundaries and good dust rims are absent but the ghost grains are still visible. Some of these grains may have been flattened during the metamorphic process but the general granular structure persisted during recrystallization and the gas, liquid and minor accessories which originally occupied the pore spaces are believed to have been trapped in their approximate positions as the quartz recrystallized. The newly formed crystals include several of the original grains and grain boundaries. Within the original grain the inclusions are unoriented, but pressure is believed to have concentrated and aligned the inclusions of the voids around the original grain.

Of the 14 slides of quartz pebbles, 12 show definite ghost grains. In the majority of these slides the ghost grains will be well-developed in some areas and less certain in others. The quartz of the remaining two slides shows a fracture mosaic which masks the ghost structure and makes its presence uncertain. Because most of the pebbles show these ghost grains their source is believed to be a metaquartzite rather than vein quartz.

A further check to distinguish metamorphic from igneous quartz was made using the type of inclusions demonstrated by Mackie (1896, p. 154). According to Mackie's law, under a magnification of 100 times or more, quartz with acicular inclusions is derived from an igneous rock and quartz with regular or crystal inclusions is derived from a metamorphic source. Irregular inclusions are common to quartz of either source and, therefore, are not definitive. Other workers (Gilligan, 1920; Tyler, 1936) have used Mackie's discovery, apparently with success, to determine the source rock for sediments. The closest use of Mackie's law to the Sharon geographically is the work of Tyler (1936) in the St. Peter sandstone of Wisconsin.

Bowen's thin sections of quartz pebbles were examined for types of inclusions. No acicular inclusions are found but all specimens have regular crystal inclusions and abundant irregularly shaped gas and liquid inclusions. If Mackie's law is valid the quartz of these pebbles was derived from metamorphic rocks.

The very high percentage of quartz in the Sharon suggests that the original source rock for the pebbles must have been either a pure metaquartzite or a quartz-rich metamorphic rock which has been subjected to several cycles of erosion with the elimination of most of the impurities.

Inclusions in quartz of the orthoquartzite are similar to those in the pebble except that a few grains have acicular inclusions. Under Mackie's law the sand grains of the orthoquartzite also would be considered of metamorphic origin with a small percentage of igneous origin.

Study of inclusions in both pebbles and sand grains of the Sharon conglomerate was only exploratory. It is hoped that a more detailed study can be undertaken which will attempt to trace the material back to source rocks in Canada and/or older clastic rocks in Ohio on the basis of heavy minerals and similar inclusions in the quartz.

Summary of Features Indicating a Northern Source

The following features imply a northern source for the Sharon conglomerate of Area B:

(1) Southerly trending conglomerate belts whose merging pattern suggests distributary streams on a deltaic or aluvial plain
(2) Cross-bedding ranging in dip from southeast to southwest
(3) Creep of top of cross-beds to give overturning to the south
(4) Gentle southerly slope of the tops of cross-beds and horizontal beds which is markedly steeper than the regional dip.
(5) Pattern of imbricate structure locally where flattened pebbles are found
(6) More abundant lenses of orthoquartzite in conglomerate in the southern part of the area, indicating more rapid changing of sedimentation conditions
(7) More abundant trough cross-bedding in the southwest indicating more rapidly changing conditions here than nearer the source to the north.
(8) Progressive decrease southward in amount of pebble along a pebble belt from 35 to 54 per cent pebble in the north to 25 to 32 per cent in the south as shown by Bowen's mechanical analyses
(9) Elimination of fragile types of odd pebbles from north to south reported by Bowen
(10) Decrease in quantity of more resistent odd pebbles from north to south reported by Bowen

Summary of Features Indicating a Sedimentary and Metamorphic Source

The following features prove that the Sharon conglomerate is the result of erosion of a mixed sedimentary and metamorphic series and that an igneous source, suggested by the erroneous name "vein quartz" for the pebbles, is of minor importance.

(1) High purity (96+% SiO_2) of the formation suggesting at least a second generation sand
(2) Both sand grains and pebbles with secondary quartz crystal faces which are characteristic of second generation sands
(3) Pebbles of sandstones, conglomerates, rotten limestones and silicified Middle Devonian fossils
(4) Several colors of quartzite pebbles with grain structure indicating at least several beds of quartzites
(5) Clear and milky quartz pebbles ("vein quartz") with ghost grain structure and regular inclusions characteristic of quartz recently derived from a metamorphic rather than an igneous terrain
(6) Most quartz grains with inclusions characteristic of quartz derived from a metaquartzite
(7) Rare quartz grains with acicular inclusions characteristic of recent igneous rock source
(8) Absence of igneous rocks, igneous rock minerals and a suite of heavy minerals typical of igneous rocks

Character of Basin of Deposition

Early ideas on the horizontal extent and vertical thickness of the Sharon conglomerate were colored by three different concepts which

gave a false picture of the shape of the Sharon deposit and its basin of deposition. First, the extreme difficulty of separating Massillon sandstone (orthoquartzite) from the orthoquartzite phase of the Sharon when the intervening Sharon shale is absent, caused many

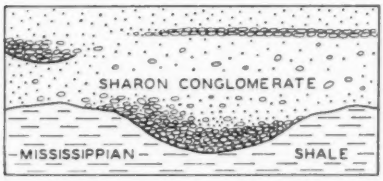

FIGURE 3.—RELATIONSHIPS OF CONGLOMERATE TO SANDSTONE IN THE SHARON

thick sections of sandstone to be classed as Sharon (Newberry, 1878, p. 137–138) when actually both Sharon and Massillon are present. Second, experience in mining the Sharon coal indicated that the coal lay in a "series of troughs" (Roy, 1875; Orton, 1884, p. 227) 150–900 feet wide trending in a southwest direction. The underlying Sharon conglomerate was also believed to be confined to troughs, and further, the conglomerate was believed to extend north of the coal basins still confined to troughs. Third, the lack of recognition of the fact that the Sharon is dominantly an orthoquartzite instead of a conglomerate and, therefore, that much of the orthoquartzite away from the conglomerate belts is Sharon rather than Massillon, supported the idea of a formation confined to long narrow depressions.

Detailed mapping of the Garrettsville quadrangle and parts of the quadrangles to the north and west has established the Sharon as a sheet-like deposit with a fairly flat top and an irregular base, caused by filling of the irregular topography of the Mississippian surface. In this sheet are belts of conglomerate with a southerly trend, which lie at any elevation but mainly in the lower part (Pl. 2, fig. 4), and which grade laterally into orthoquartzite (Fig. 3). This whole sheet, as shown by drill records and sections, thickens westward from the west edge of the Grand River Valley and formed in a large basin, the western extent of which is not known.

ENVIRONMENT OF DEPOSITION

Early ideas on the environment of deposition were diversified. As stated previously Newberry (1873, p. 213) considered the Sharon to be a glacial deposit dumped in a series of long narrow troughs; Stout (1916, p. 443), a sea current deposit restricted "to narrow passages"; and Lamb (1911, p. 105), a stream deposit.

Cross-bedding and high degree of sorting are so prominent in any large outcrop of Sharon that geologists today are certain that the Sharon is a water-laid deposit. The specific kind of water-laid environment is not as easily agreed upon. Some have suggested that the Sharon is a beach deposit. Thompson (1937) determined criteria for recognition of ancient bars, dunes and beaches from a study of modern beaches. Many features found in modern beaches are found in the Sharon on a larger scale. Most significant, however, are Thompson's conclusions (1937, p. 744) that ancient beaches are characteristically thin, generally less than 40 feet thick, that they rapidly lens and interstratify with marine or continental sediments, and that they are scarce in the geologic record. It seems certain that the extensive deposit of Sharon cannot be a beach deposit or even a transgressive beach deposit, although local areas may have some beach-deposited material.

At one locality a buried scarp has been found and deposits nearby suggest characteristics of upper foreshore beaches overlain by lower foreshore beaches (Thompson, 1937, p. 726–740).

It has been suggested that the Sharon may be a series of offshore bars, bars, and spits. Characteristically these forms have complicated cross-stratification, while the cross-stratification of the Sharon is consistent. These deposits, like beaches, are of such limited areal extent that the Sharon can not be considered to be dominantly a series of bars.

The third possible environment of deposition is an alluvial or deltaic plain. Deposition in this environment most readily explains these characteristics which have already been discussed: (1) widespread thick deposit with

irregular base and relatively flat top, (2) radial distribution of the conglomerate belts, (3) long lateral extent of some conglomerate belts, (4) uniformly high degree of sorting in individual beds, (5) consistent widespread simple cross-stratification, (6) widespread initial slopes, and (7) lag gravels.

Further work may define the detailed extent and character of the alluvial or deltaic plain and its source. The present concept is that of a highland in Canada which was composed of a Pre-pennsylvanian sequence of well-cemented quartzites, sandstones, conglomerates and limestones as dominant types. As this highland was eroded and the material transported to the south, rounded quartz pebbles were formed from the quartzites and the pebbles of the conglomerate, and abundant sand was produced from the sandstone and the matrix of the conglomerate. Most of the limestone went into solution, but some remnants persisted especially where silicified; among this silicified material were the fossils.

This material moved southward on an appreciable slope and was deposited in a broad basin near sea level, forming an alluvial or deltaic plain on which the distributary channels shifted position frequently. As these channels shifted, the load built forward in prominent cross-beds and layer after layer of cross-bedded orthoquartzite was deposited. At times increased current eroded the tops of these cross-beds, and as the velocity decreased again, nearly horizontal beds were deposited, or sheet wash beyond the active channels spread a similar horizontal layer.

Normally the stream gradient was such that the sand was carried and the pebbles rolled along the bottom to give a mixture of sand with scattered pebbles. Local increase in velocity in a channel sorted out the sand, and the pebbles concentrated in the deeper part of the channel, where they later came to form the conglomerate belts. Increase of velocity on the main plain, away from the channel, also caused erosion, which moved the sand farther south and concentrated the gravel at a lower level in a horizontal layer to form the lag-gravel sheets.

Comparison with Area A

Mapping in Area A has proceeded far enough to give a fair concept of the Sharon, and since this information is supported by that in the literature (White, 1880), it is evident that the Sharon of Areas A and B are similar. This deposit is mainly orthoquartzite with one known southerly trending conglomerate belt. The formation is thinner and the pebbles are smaller than those of Area B, which suggests that this area was farther from the source or had smaller streams to distribute the material.

Area A is separated from Area B by the broad Grand River lowland cut in Mississippian formations. Considering that the Sharon thins toward the east in Area B, and that the Sharon-Mississippian contact rises in this area, the writer (1950) believes that the present Grand River Valley represents a Mississippian highland which separated the basin of Area A from Area B.

That Area A is distinct from Area B is confirmed by the heavy mineral studies of Rittenhouse (1946, p. 1195), who says that the easternmost Sharon (Area A) and the Olean had the same source which was different from that of most of the Sharon (Area B) of northeastern Ohio.

Comparison with Area C

The writer and two students have mapped three townships in Area C (Jessup, D. E., 1951, thesis, Ohio State University, 97 p.; Weiss, R. M., 1951, thesis, Ohio State University, 101 p.). This work and reconnaissance indicate that the Sharon in Area C is similar to that of A and B. Similarity lies in the presence of pure orthoquartzite with scattered pebbles and linear pebble belts, well-developed cross-bedding, silicified tabulate corals, and a few pebbles derived from sedimentary rocks. The thickness of the Sharon in Area C is similar to that in Area A; Stout (1916, p. 42) reports a maximum of 200 feet. One conglomerate channel trending in a northwest-southeast direction has been followed for 9 miles and another for 6 miles. Both the northwest-southeast trend of the channels and the west

to north dip direction of the cross-beds indicate a southeastern source for the Sharon of Area C. Rittenhouse (personal communication) feels, however, that Areas B and C had a common source on the basis of "(1) similarity in roundness in heavy minerals in both areas, (2) similarity in chromite concentration in both areas (the relatively high chromite content in the Sharon is unique to that formation), and (3) the presence of Devonian fossils in Area C." These similarities do indicate that the original sources for the two deposits must have been very similar but not necessarily that they were the same.

Work by many geologists (Morse, 1910; Stout, 1916, 1918; Conrey, 1921; Meyers, T. R., 1929, thesis, Ohio State University, 85 p.; White, 1949; Merrill, W., 1950, dissertation, Ohio State University, 444 p.; Flint, 1951; Hall, J. F., 1951, dissertation, Ohio State University, 218 p.; and Lamborn, 1954) has helped to prove a Mississippian highland in central Ohio. In Perry, southwestern Muskingum, eastern Licking, and northern Hocking counties the existence of the highland is shown by widespread areas of the youngest Mississippian formation, the Maxville limestone. In these same counties the oldest Pennsylvanian, Sharon conglomerate, is found only in the lowest valleys, and higher Pennsylvanian beds commonly overlie the Mississippian. In Coshocton and Holmes counties the highland is indicated by the complete lack of Sharon conglomerate. North and south of this highland lie the basins in which the Sharon conglomerate of Areas B and C were deposited.

Conclusion

Study of the Sharon conglomerate of northeastern and southern Ohio has modified older concepts of the character and origin of the Sharon. The writer has previously reported on some of these discoveries (1947; 1950); this paper summarizes the newer ideas on character and origin of this economically important pure orthoquartzite and conglomerate (96+ per cent SiO_2).

Mapping has established three distinct Sharon areas in which the formation lies in basins on the original rolling Pre-pennsylvanian surface. The base of the Sharon is, therefore, irregular but the top is fairly regular. Dip of cross-bedding throughout the formation indicates deposition from rapidly flowing streams whose source was to the north in northern Ohio and to the southeast in southern Ohio. These source locations are confirmed by conglomerate belts with a southerly trend in northern Ohio and a northwesterly trend in southern Ohio. The radial pattern of the conglomerate belts suggests deposition on a deltaic or alluvial plain. In Area B dip of initial slope, direction of over-turning of tops of cross-beds, imbricate structure, and north-south changes in mechanical analyses, chemical analyses, and other sedimentary features also indicate a northern source.

In all areas the conglomerate belts contain shale lenses and masses which are believed to be quiet water deposits formed away from the channels of the stream, and later covered or eroded and moved to become clay mass pebbles. Well-sorted thin conglomerate beds from one pebble to several feet thick are found in the orthoquartzite and conglomerate. They are believed to be reworked topset beds or older cross-beds and are called lag gravels.

Pebbles of sedimentary rocks and Middle Devonian fossils as pebbles indicate that the source rocks for the Sharon must have been in part sedimentary. In northeastern Ohio (Area B) discovery of ghost grain structure and inclusions in the quartz characteristic of metamorphic quartz indicates that the so-called "vein quartz" pebbles were not derived from an igneous source but are metamorphic in origin. The writer, therefore, feels that the northern source for the detritus of the Sharon was a mixed sedimentary and metamorphic terrain instead of igneous as was formerly suggested.

References Cited

Bowen, C. H., 1953, Petrology and economic geology of the Sharon conglomerate in Geauga and Portage counties, Ohio: Ohio State Univ. Eng. Expt. Sta. Bull., no. 153, 58 p.

Bownocker, J. A., 1921, Steal molding sands in Ohio: Ohio Jour. Sci., v. 21, no. 8, p. 249–266.

—— 1926, Glass sands in Ohio: Ohio Jour. Sci., v. 26, no. 1, p. 25–41.

Conrey, G. W., 1921, Geology of Wayne County: Geol. Survey Ohio Bull. 24, 155 p.

Clements, J. M., and Smyth, H. L., 1899, The Crystal Falls iron-bearing district of Michigan: U. S. Geol. Survey Mon. 36, 512 p.

DeWitt, W. Jr., 1946, The stratigraphic relationship of the Berea, Corry, and Cussewago sandstones in northeastern Ohio and northwestern Pennsylvania: U. S. Geol. Survey Oil and Gas Inv., Prelim. chart 21.

Flint, N. K., 1951, The geology of Perry County, Ohio: Geol. Survey Ohio Bull. 48, 4th ser., 234 p.

Fuller, J. O., 1947, Sharon conglomerate, a source of high-silica raw material: Ohio State Univ. Eng. Sta. News, v. 19, no. 2, p. 48–55.

—— 1950a, Northern Middle Devonian sedimentary source for the Sharon Conglomerate of northeastern Ohio (Abstract): Geol. Soc. America Bull., v. 61, p. 1463–1464.

—— 1950b, New discoveries concerning the Sharon conglomerate in northeastern Ohio (Abstract): Geol. Soc. America Bull., v. 61, p. 1567.

Gilligan, Albert, 1919, The petrography of the Millstone grit in Yorkshire: Quart. Jour. Geol. Soc. London, v. 75, p. 251–294.

Inman, D. L., 1949, Sorting of sediments in the light of fluid mechanics: Jour. Sed. Petrology, v. 19, p. 51–70.

Lamb, G. F., 1911, The Mississippian-Pennsylvanian unconformity and the Sharon conglomerate: Jour. Geology, v. 19, p. 104–109.

Lamborn, R. E., 1954, Geology of Coshocton County: Geol. Survey Ohio Bull. 53, 245 p.

Lamey, C. A., 1935, The Palmer gneiss: Geol. Soc. America Bull., v. 46, p. 1137–1162.

Mackie, William, 1896, The sands and sandstones of eastern Moray: Edinburgh Geol. Soc. Trans., v. 7, p. 148–172.

McKee, E. D., and Weir, G. H., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geol. Soc. America Bull., v. 64, p. 381–390.

Morse, W. C., 1910, The Maxville limestone: Geol. Survey Ohio Bull. 13, 4th ser., 128 p.

Newberry, J. S., 1873, Report of the Geological Survey of Ohio: Geol. Survey Ohio, v. 1, pt. 1, 680 p.

—— 1873, Report of Geological Survey of Ohio: Geol. Survey Ohio, v. 3, pt. 1, 958 p.

Orton, Edward, 1884, Report of the Geological Survey of Ohio: Geol. Survey Ohio, v. 5, Econ. Geology, 1124 p.

Rittenhouse, Gordon, 1946, Grain roundness—a valuable geologic tool: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 1192–1197.

Roy, Andrew, 1875–76, The Mahoning Valley coal region: Am. Inst. Min. Engineers Trans., v. 4, p. 188–190.

Russell, R. D., 1939, Effects of transportation on sedimentary particles *in* Trask, P. D., *et al.*, Recent marine sediments, a symposium: Am. Assoc. Petroleum Geologists, p. 32–47.

Stout, Wilber, 1916, Geology of southern Ohio: Geol. Survey Ohio Bull. 20, 4th ser., 723 p.

—— 1918, Geology of Muskingum County: Geol. Survey Ohio Bull. 21, 4th ser., 351 p.

Thompson, W. O., 1937, Original structures of beaches, bars, and dunes: Geol. Soc. America Bull., v. 48, p. 723–752.

Tyler, S. A., 1936, Heavy minerals of the St. Peter sandstone in Wisconsin: Jour. Sed. Petrology, v. 6, p. 55–84.

White, G. W., 1949, Geology of Holmes County: Geol. Survey Ohio Bull. 47, 4th ser., 373 p.

White, I. C., 1880, The geology of Mercer County: 2nd Geol. Survey Pa., Rept. Progress 3Q., 233 p.

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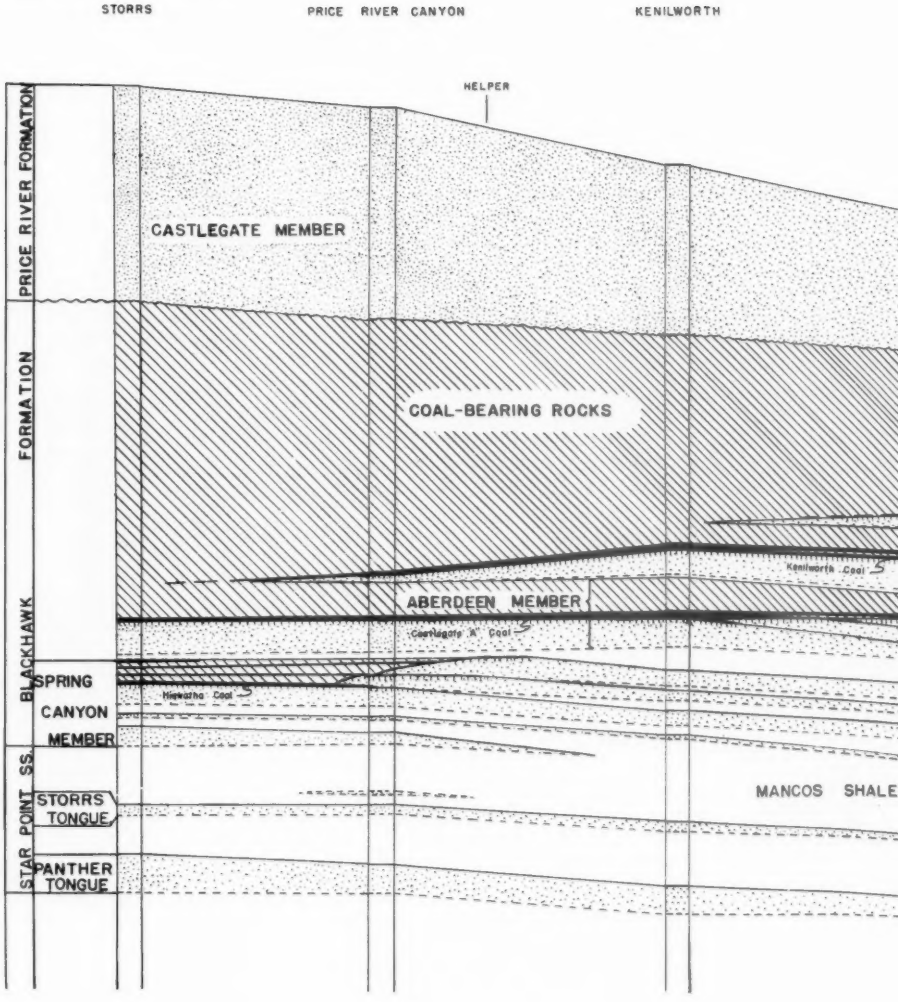
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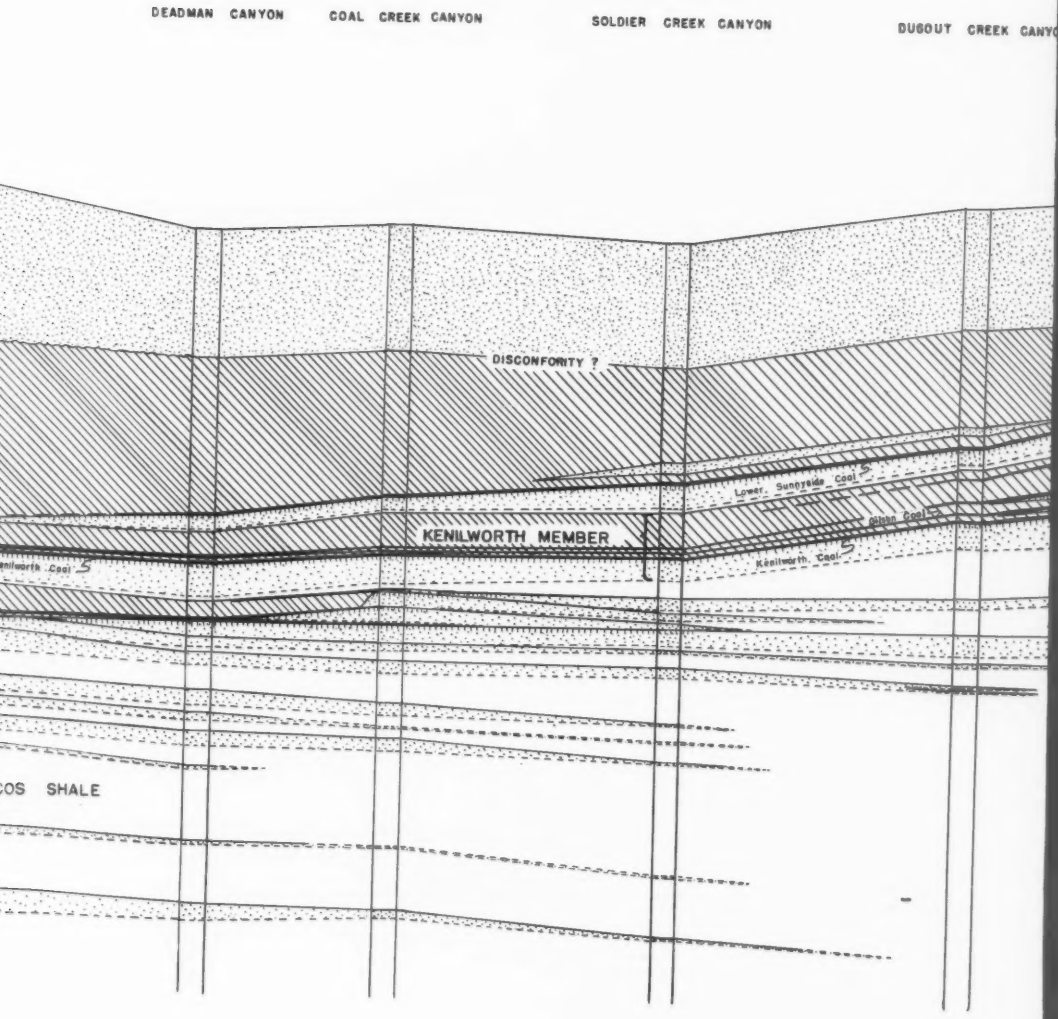
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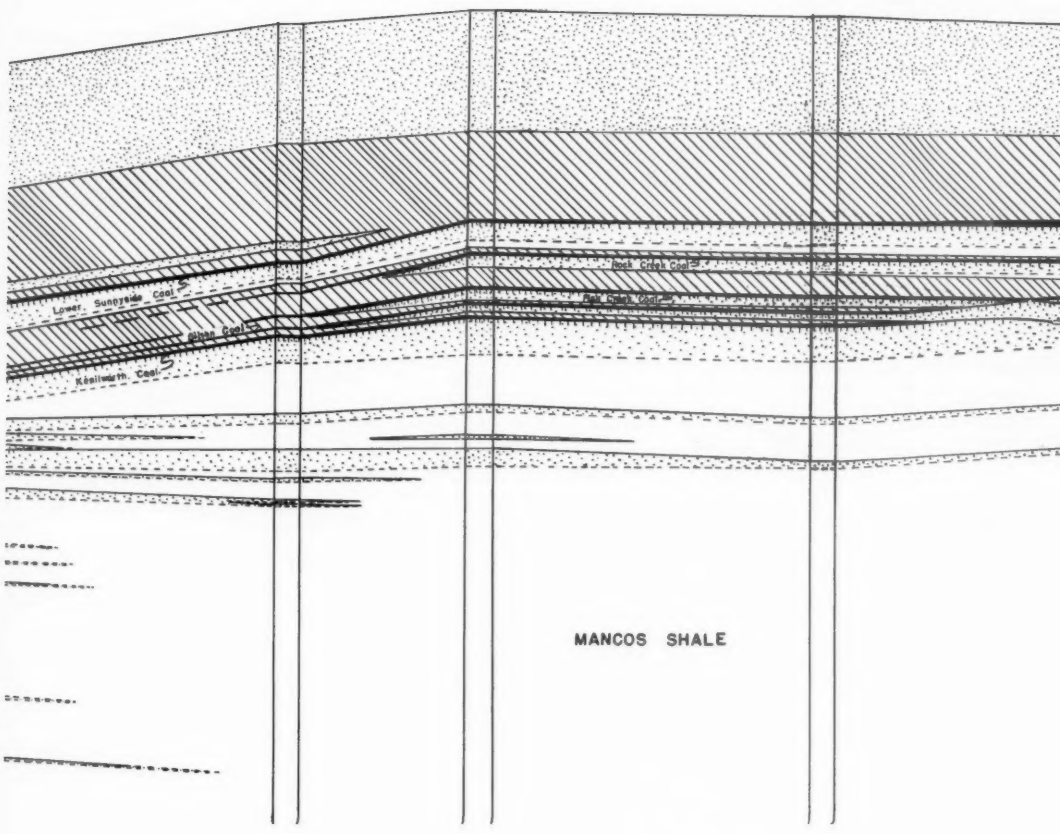

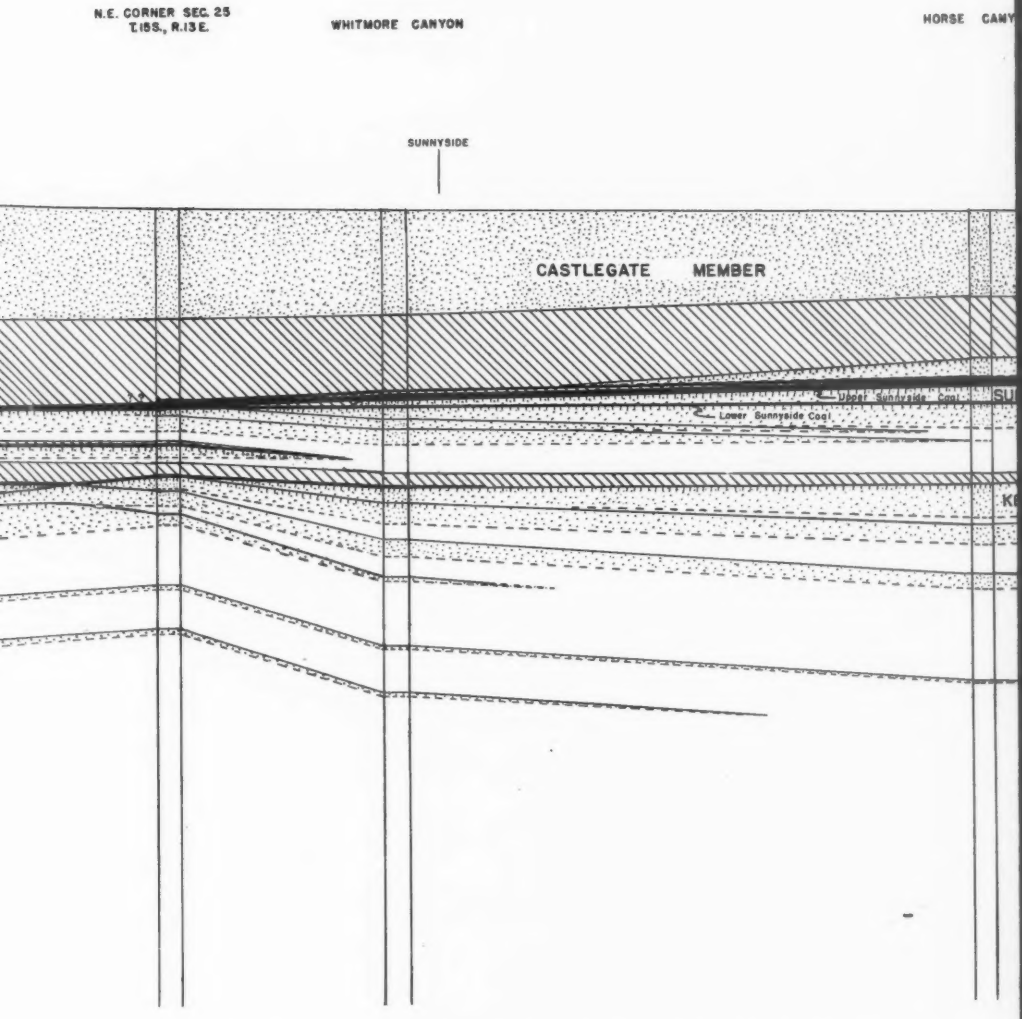

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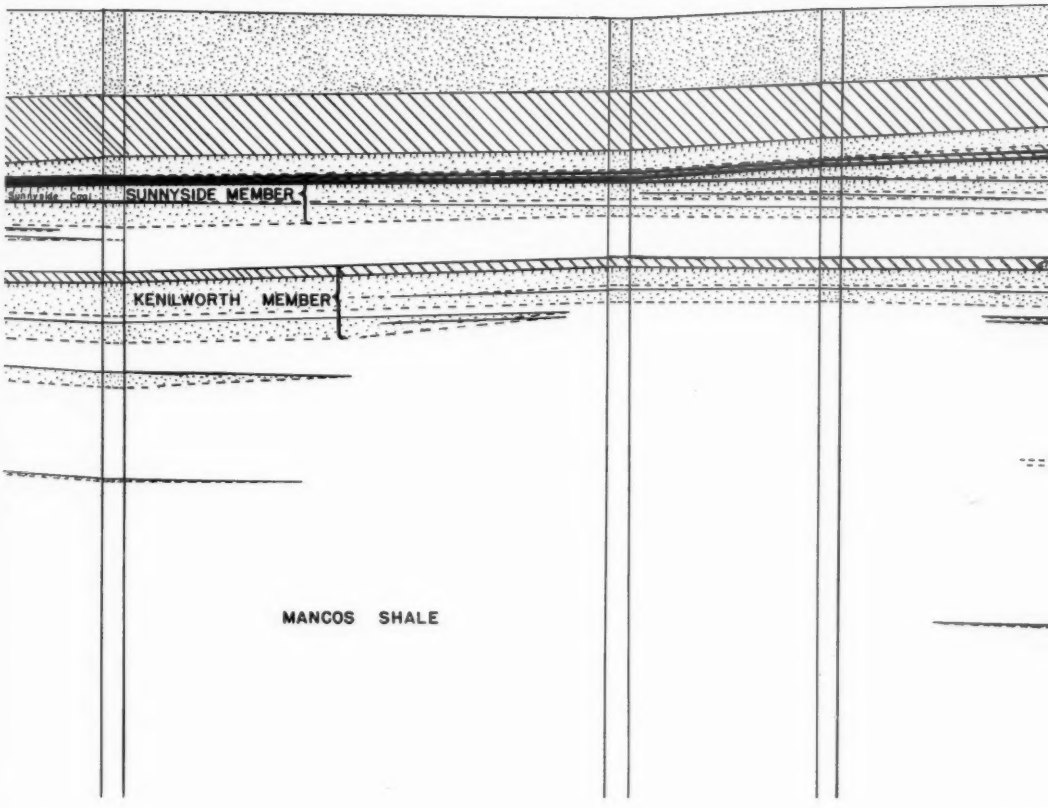

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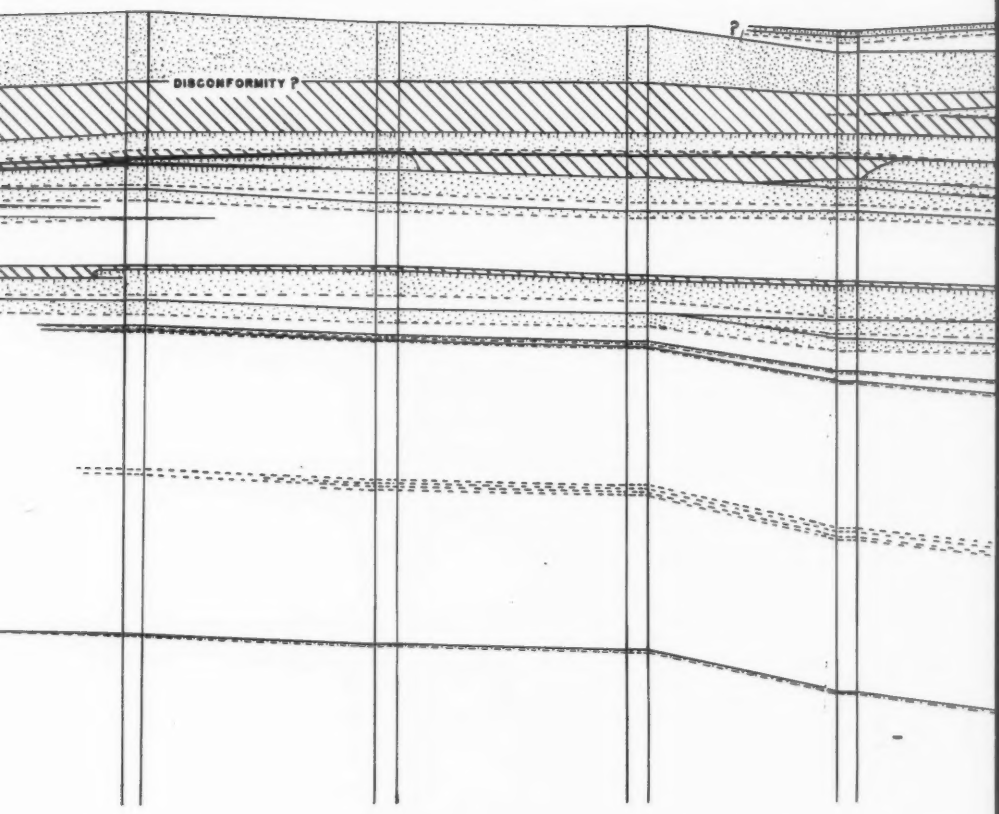

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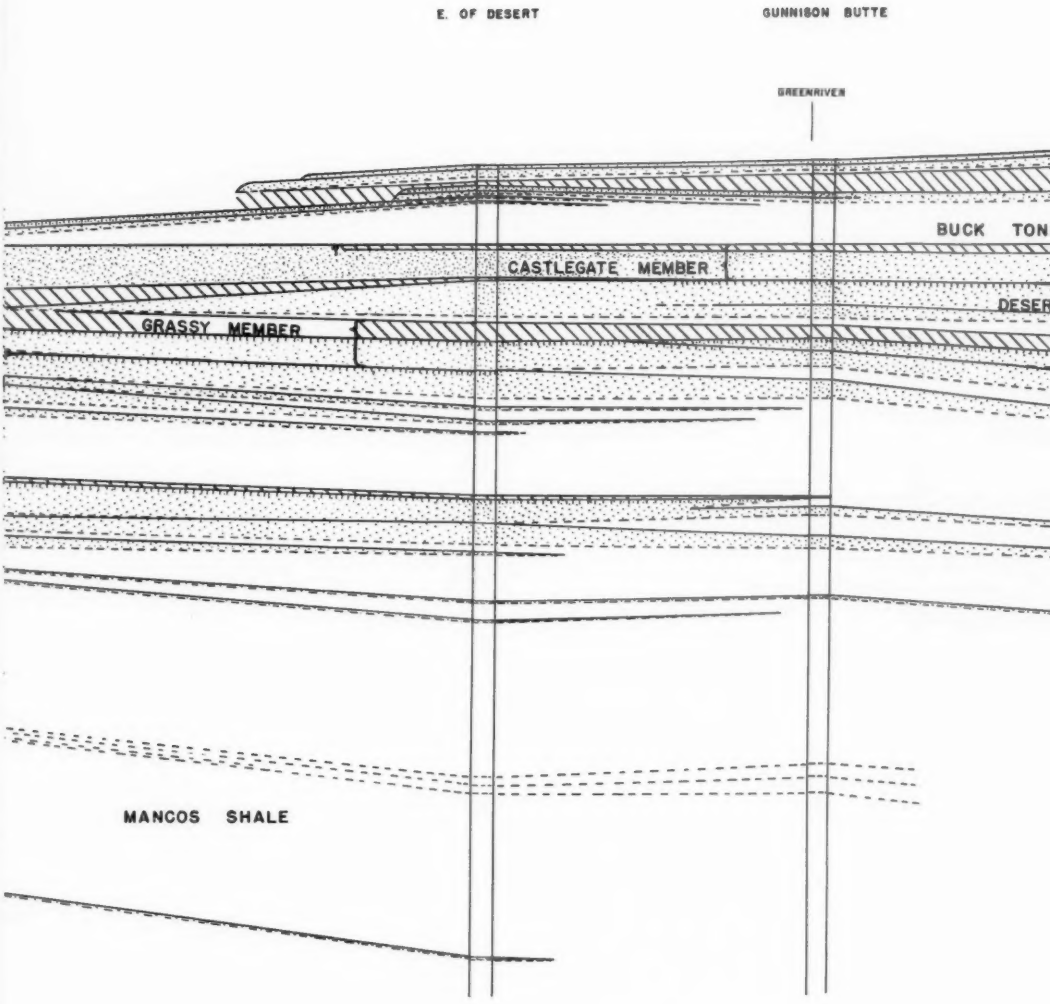

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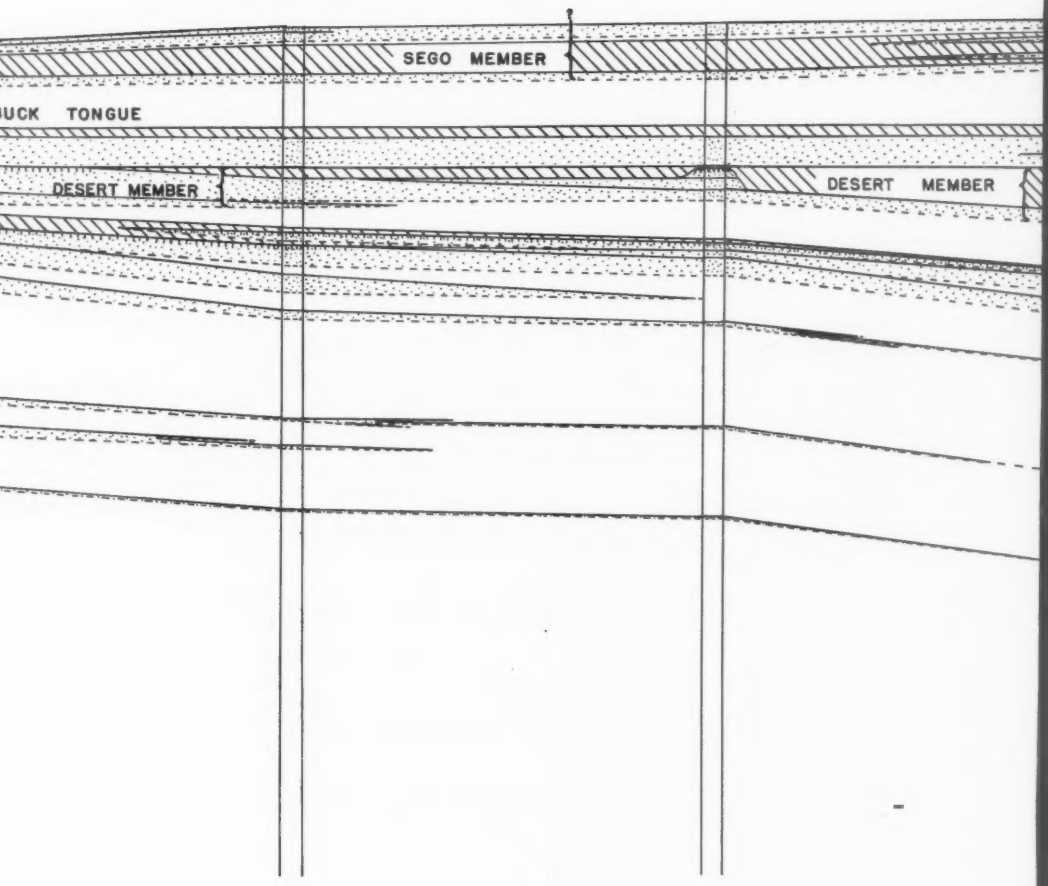
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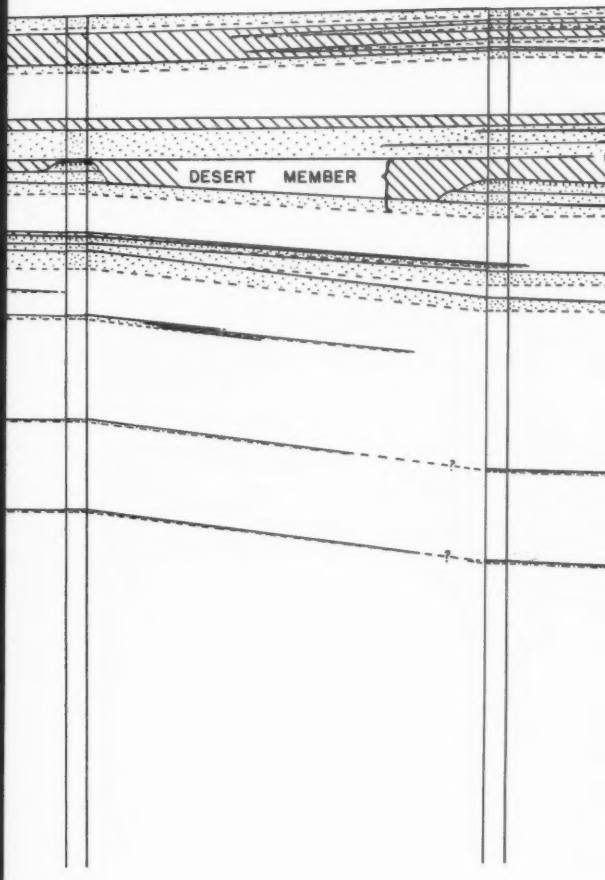

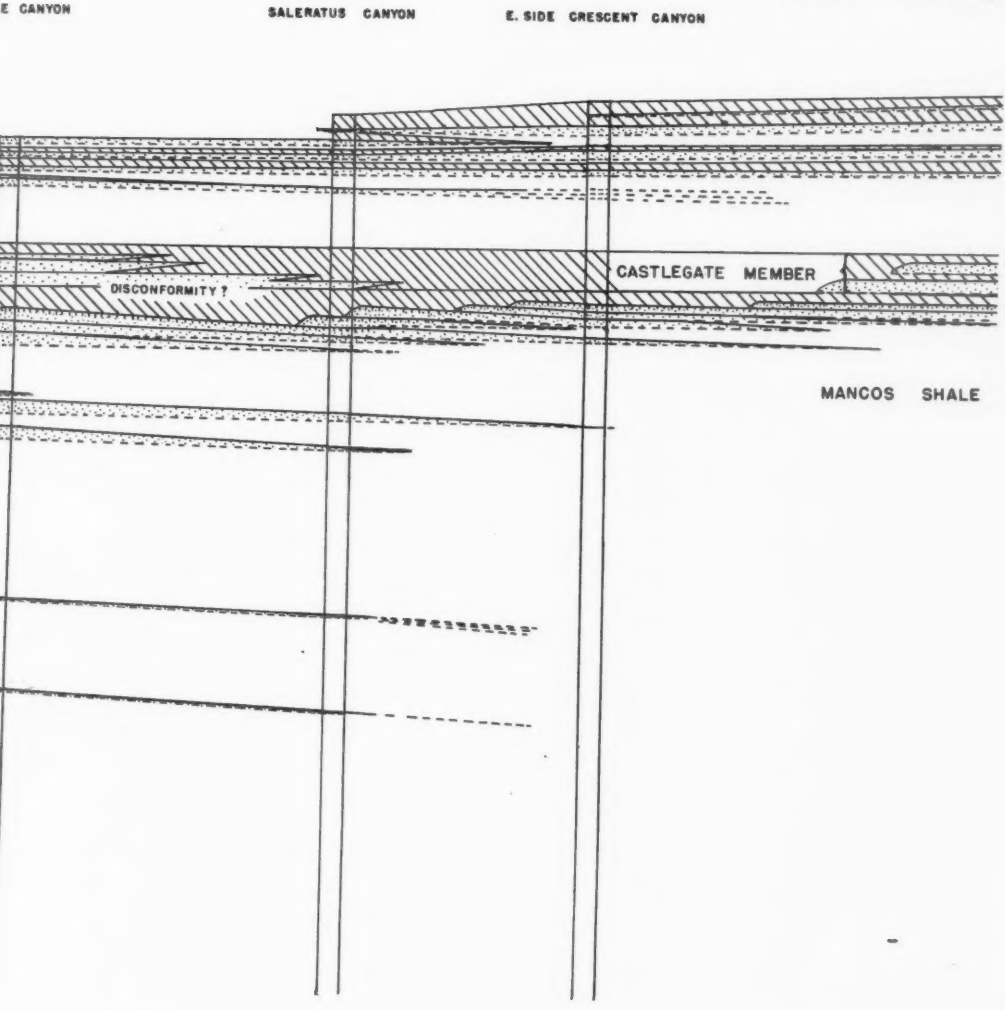
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VOL. 66, PP. 177–202, 4 FIGS., 3 PLS. FEBRUARY 1955

SEDIMENTARY FACIES AND INTERTONGUING IN THE UPPER CRETACEOUS OF THE BOOK CLIFFS, UTAH-COLORADO

By Robert G. Young

Abstract

The Upper Cretaceous of the Book Cliffs in central Utah and western Colorado exhibits intricate lateral and vertical intertonguing of marine and nonmarine facies. Excellent exposures permit detailed observations and facilitate tracing of individual units. Dominantly continental deposits of the Star Point, Blackhawk, and Price River formations pass eastward into lagoonal deposits formed behind offshore bar sandstones, which interfinger eastward with the marine Mancos shale.

The Star Point sandstone consists of two littoral marine sandstone tongues. Overlying the Star Point is the Blackhawk formation consisting of six members separated by thin tongues of Mancos shale, which grade upward into the overlying littoral marine sandstones. Each marine shale tongue rests with a slight disconformity on the underlying member. Where the littoral marine sandstones are absent the rocks cannot be so subdivided. The Price River formation which rests disconformably on the Blackhawk consists of two facies—the Farrer or barren facies and the Neslen or coal-bearing facies which is divided into five cyclic members similar to those of the Blackhawk.

The intricate intertonguing of these deposits is interpreted to be the result of deposition in a shallow basin in which there were long periods of relative stability separated by sharp pulses of subsidence. Thick peat beds formed during periods of quiet coincide with the tops of offshore bars. Sharp subsidences preceded formation of the basal sandstone tongues; lesser pulses preceded formation of the offshore bars.

A generalized cycle of four units can be recognized in these deposits: (1) basal marine shale, (2) littoral marine sandstone, (3) lagoonal rocks, (4) coal.

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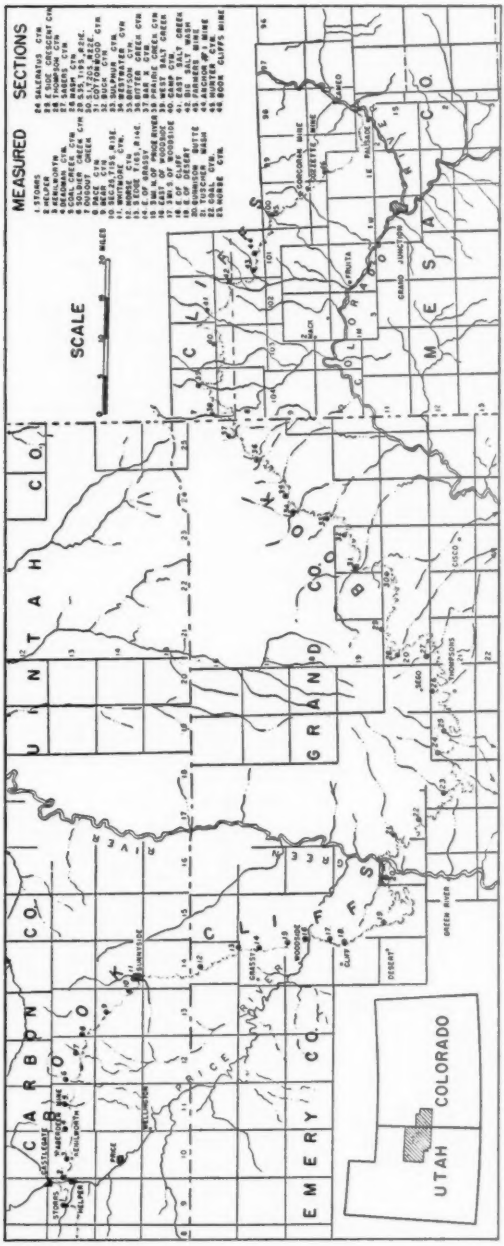

FIGURE 1.—INDEX MAP

Introduction

The Book Cliffs of central Utah and western Colorado present unexcelled exposures of Upper Cretaceous continental, littoral marine, and marine strata which can be traced continuously for more than 200 miles. These rocks form nearly vertical, barren cliffs in which the stratigraphic relations can be readily ascertained. Many sections of the steep cliffs are inaccessible, and study of these strata at close range is extremely hazardous.

The area studied extends eastward from the eastern margin of the Wasatch Plateau in central Utah along the Book Cliffs, which form the northern rim of the Colorado Plateaus in eastern Utah, to Grand Junction, Colorado (Fig. 1) about 220 miles. The width of outcrop studied ranges from a fraction of a mile to about 3 miles. Along the entire exposure structure of the rocks remains simple, with the entire sequence dipping a few degrees north off the northern and eastern flanks of the San Rafael Swell and off the northern and northwestern flanks of the Uncompahgre uplift with only a few faults and folds.

The gross nature of the intertonguing strata had already been determined by Spieker and Reeside (1925), Clark (1928), Erdmann (1934), and D. J. Fisher (1936). For more detailed study the writer spent 8 months—the summers of 1950 and 1951—in the field in the Book Cliffs area measuring more than 55 sections at accessible points with hand level and alidade. These sections were correlated in detail by tracing individual beds from one section to another. Five detailed sections are presented in the writer's dissertation (Young, R. G., 1952, unpublished PhD dissertation, The Ohio State University, Columbus, Ohio, 133 p.).

The rock units studied in detail are the Star Point sandstone, the Blackhawk formation, and the Castlegate and Sego members of the Price River formation, and the corresponding units of the Mancos shale.

Acknowledgments

A research fellowship from The Ohio State University and additional assistance from the geology department of The Ohio State University made this project possible. The writer is deeply indebted to Edmund M. Spieker for suggesting the problem, giving much needed assistance and counsel during the study, and constructive criticism of the manuscript. Gratitude is also expressed to Richard P. Goldthwait and Charles H. Summerson of The Ohio State University, and to Edwin D. McKee of the United States Geological Survey for helpful suggestions concerning various problems.

Thanks are due Bert J. Silliman of Green River, Utah, and Emery Nielson of Fruita, Colorado, for supplying much regional data and living quarters. The writer is also grateful to his wife, Joan, who assisted him in the field and in the preparation of the manuscript.

General Features

The Upper Cretaceous rocks exposed in the Book Cliffs consist of sandstones, thin shales, and one or more coal-bearing units, all of which intertongue eastward with marine Mancos shale (Fig. 2). The units involved are the Star Point sandstone and the Blackhawk and Price River formations or their equivalents. As a result of this eastward intertonguing, the contact between the littoral marine sandstones and the underlying Mancos shale rises 2700 feet stratigraphically from the western end to the eastern end of the Book Cliffs (Spieker, 1949, p. 67). Since the sandstones form the Book Cliffs, the strata exposed in the eastern portions of the cliffs are younger than those in the cliffs to the west.

The eastward intertonguing of the sandstones with shale reflects movement of the strand line of the Upper Cretaceous sea. Since the Book Cliffs trend slightly southeast from the Wasatch Plateau, they present a cross section nearly at right angles to the Upper Cretaceous shore line, which in this region had a northeast trend. As the basin filled, the strand line moved eastward. Minor oscillations of the sea are recorded in the zigzag path of the strand line which is represented by the littoral marine sandstone tongues.

Classification and Nomenclature

The problem of classification and nomenclature is a difficult one. At present there is

no standard procedure for dealing with intertonguing deposits such as those of the Book Cliffs. This is due largely to the fact that records, such as those of the Book Cliffs, are rare and also to the fact that our present stratigraphic framework is inadequate to deal with such deposits.

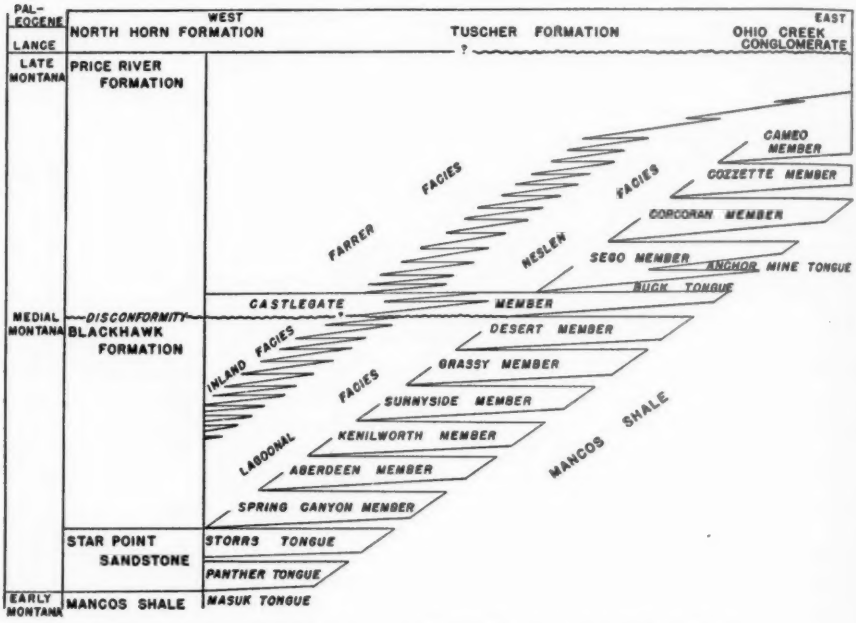

FIGURE 2.—DIAGRAMMATIC SECTION OF ROCKS EXPOSED IN THE BOOK CLIFFS

Subdivision and classification of these rocks by earlier workers has not been consistent. The Spring Canyon, Blackhawk, and Price River formations were defined by Spieker and Reeside (1925, p. 442, 443, 445) and the North Horn formation by Spieker (1946), p. 132). Clark (1928, p. 17) gave the names Panther, Storrs, and Spring Canyon to the three main sandstone tongues of the Star Point, and applied the name Aberdeen sandstone member to a prominent sandstone near the base of the Blackhawk in the western Book Cliffs, noting that it split eastward into two tongues separated by a tongue of Mancos shale. He also named the lower sandy member of the Price River the Castlegate sandstone.

D. J. Fisher (1936, p. 12) divided the Blackhawk into Lower, Middle, and Upper sandstone members and included the intervening Mancos shale tongues (Middle and Upper shale members) in the Blackhawk. He named the tongue of Mancos shale lying above the Castlegate sandstone member the Buck tongue, applied the name Sego sandstone member to the sandstone resting on the Buck tongue, and divided the overlying, dominantly nonmarine rocks of the Price River into the Nelsen coal-bearing member and the Farrer noncoal-bearing member.

Erdmann (1934) did not use the term Price River but established the Mt. Garfield and Hunter Canyon formations, roughly corresponding to the Neslen and Farrer members of Fisher. He recognized an upper and a lower tongue of the Sego sandstone and named the intervening tongue of Mancos shale the Anchor Mine tongue. He retained the name Rollins sandstone member as applied by Lee (1912) to a massive littoral marine sandstone near the top of the Mt. Garfield formation in Grand Mesa.

The writer has utilized the cyclic character of these rocks in grouping the littoral marine sandstone tongues with coal-bearing rocks and excluding the marine shale tongues. Each member consists of a basal white-capped sandstone and the overyling lagoonal and offshore bar deposits. Between each pair of members is a thin tongue of Mancos shale which diminishes westward to a feather edge where the littoral marine sandstone disappears into the coal measures. Each shale wedge rests with slight disconformity on the coal-bearing rocks of the next lower member, and each grades upward into the littoral marine sandstone at the base of the next higher member.

Actually these sandstones are no more closely related to the coal measures than to the marine shales—probably less so physically since the sandstones grade into the shales. However, they are here united with the coal measures since both are dominantly landward deposits.

Another logical subdivision would be one which separates the littoral marine sandstones (as stratigraphic units) from the coal-bearing rocks on the one hand and from the marine shales on the other hand. However, such a subdivision would probably necessitate the naming of numerous minor sandstone tongues and might be confusing.

A cyclic division is a valid means of subdividing these dominantly nonmarine rocks, since each member appears to represent a complete cycle of deposition and each is separated above and below by disconformities. The shale tongues are the basal units of the cycles but are still stratigraphically classified as named or unnamed tongues of the Mancos. The members of the Blackhawk and Price River are also tongues, dual or multiple, in contrast to the unitary shale tongues.

Wheeler and Mallory (1953) proposed that in the Book Cliffs the marine shale tongues be included with the littoral marine sandstones and coal-bearing rocks rather than with the main body of Mancos shale. They applied the term "Arbitrary Cut-off" to arbitrarily chosen boundaries between laterally adjoining units, the plan partly followed by D. J. Fisher (1936) in subdividing these deposits in eastern Utah. Fisher recognized the Lower, Middle, and Upper sandstone members of the Blackhawk (his terminology) separated by the Middle and Upper shale tongues of the Blackhawk. These shale members are tongues of the Mancos and in no way differ from the next higher tongue of Mancos which he named the Buck tongue.

This method of subdivision does not seem to be acceptable because, as pointed out by A. G. Fisher (1954, p. 927), it tends to obscure the intertonguing. It does not seem logical to separate these marine shale tongues from the main body of the Mancos shale, of which they are integral parts, and to place them with the dominantly continental beds to which they are much less closely related.

A facies scheme of classification such as proposed by Caster (1934) for the Devonian deposits of Pennsylvania might well be employed here. However, because of the unwieldy terminology his scheme has not been widely accepted.

Mancos Shale

The Mancos shale, a very widespread unit in western Colorado, northwestern New Mexico, northeastern Arizona, and eastern Utah, was named by Cross (1899, p. 4) from exposures near the town of Mancos and along the Mancos River valley in southwestern Colorado. The soft shale forms a broad band along the sloping base of the Book Cliffs and the valleys and terraces south of the cliffs.

As exposed in the cliffs the Mancos is drab, gray, or slightly bluish; in many places, wash from overlying sandstones has coated the outcrops buff to gray. Near the top are a few thin lenses of calcareous sandstone, limestone, and a few concretionary beds. In central Utah, where it is separated from the lower members by the Emery sandstone, this upper portion of the Mancos is called the Masuk member, but to the east it is not easily separable. It ranges from about 5000 feet thick in central Utah to about 4000 feet in western Colorado. However, the exposed portion rarely exceeds 600 or 700 feet.

The upper part of the Mancos shale grades into the overlying littoral marine sandstone tongues, with which it interfingers (Pl. 1, fig. 1). These tongues of sandstone thin eastward and gradually grade into the shale. Above

each sandstone tongue is a tongue of Mancos shale thinning westward. The two largest shale tongues, which all result from small-scale marine transgressions, are the Buck and Anchor Mine tongues. This interfingering results in a stratigraphic rise of the upper boundary toward the east, making the upper part of the Mancos increasingly younger from west to east. For instance, near Helper, Utah, it is early Montana; near Green River it is medial Montana; and in western Colorado it is mostly late Montana. The base of the shale is about the same age throughout. The 4000 feet in western Colorado, therefore, represents a much longer period of geologic time than does the 5000 feet of central Utah. Since there is no evidence of erosion, time lines in the Mancos must converge strongly toward the east.

The Mancos shale was deposited as muds in the shallow water of the Late Cretaceous sea beyond the sand-mud transition line. Convergence of time lines indicates that most of the mud was deposited near the shore, resulting in differential sedimentation. Spieker (1949, p. 68) remarks on the unexplainable scarcity of limestone in marine Upper Cretaceous rocks of the western states. There are a few stringers of limestone and some ironstone concretions, but no great volume of calcareous material, and probably conditions for limestone deposition were not favorable in the shallow, muddy waters of the Late Cretaceous seas in the Rocky Mountain region.

The Buck tongue of the Mancos, above the Castlegate member and below the Sego member, was named by D. J. Fisher (1936, p. 15). From its western edge near Woodside the tongue thickens eastward to a maximum of 380 feet at West Salt Creek, where it passes into the main body of the Mancos.

Erdman (1934, p. 36) named the Anchor Mine shale tongue, which separates the upper and lower sandstone tongues of the Sego member. In places this shale actually splits into two tongues that separate three sandstone tongues of the Sego. It appears near Saleratus Canyon and reaches a thickness of 160 feet at Big Salt Wash.

Star Point Sandstone

The Star Point sandstone was named by Spieker and Reeside (1925, p. 442) for a prominent headland of the Wasatch Plateau in Carbon County, Utah, about 3 miles west of the southwest corner of the Castlegate quadrangle. Clark (1928, p. 116) used this unit as the basal unit of the "Mesaverde Group", a term no longer used in the Book Cliffs.

As originally defined, the base of this unit was the base of the Panther sandstone, and the upper boundary was the contact of the Spring Canyon sandstone with the overlying coal-bearing rocks of the Blackhawk formation. However, the Star Point sandstone of the western Book Cliffs in the area to the north and east of Spring Canyon is redefined, keeping the base as the base of the Panther sandstone, but placing the upper boundary at the top of the Storrs sandstone. Change of the upper boundary is impractical to the west and south in the Wasatch Plateau where the Spring Canyon, Storrs, and Panther sandstones unite to form a single massive littoral marine tongue (Spieker, 1931, p. 24); however, the shift is necessary in the Book Cliffs to conform with the writer's scheme of classification.

The Star Point is present only in the western part of the Book Cliffs, where a tongue of sandy Mancos shale about 200 feet thick separates the Panther and Storrs sandstone tongues. Fossils collected by Clark and Spieker date the Star Point as medial Montana.

The Panther sandstone tongue was named by Clark (1928, p. 17) for exposures in Panther Canyon where it is about 70 feet thick. It has a sharply defined top, is a conspicuous cliff-former near Helper, Utah, and can be traced from the front of the Wasatch Plateau, where it is 125 feet thick, to a short distance east of Soldier Canyon at which point it disappears in the Mancos shale. In Soldier Canyon its position is indicated by a band of impure limestone concretions and thin lenses of sandstone. The coarseness and degree of cementation decrease from west to east. Marine cross-bedding is characteristic, and in places the upper surface has undergone some channeling. The lower part grades downward from sandy shale into shale, and the base of the sandstone is placed at the point where sandstone becomes dominant over shale. Individual sandstone layers in the tongue range from 1 inch to 10 feet thick, and in places where the beach was very steep one can see these individual beds

pass downward through the sandstone into the shale below (Spieker, 1949, p. 62).

About 120 feet above the top of the Panther tongue is the Storrs sandstone tongue, named for exposures at the town of Storrs in Spring Canyon (Clark, 1928, p. 17). The Storrs, like the Panther, has a sharply defined top and grades downward from thick-bedded, medium-grained sandstone into sandy shale and shale. It is about 30 feet thick at Storrs and thins to the east, disappearing about 1 mile east of Kenilworth, Utah. It is separated from the underlying Panther and overlying Spring Canyon sandstone tongues by westward-pointing tongues of Mancos shale.

Blackhawk Formation

Spieker and Reeside (1925, p. 443) applied the name Blackhawk formation to the coal-bearing rocks exposed in the western part of the Book Cliffs and in the eastern front of the Wasatch Plateau. Consisting of about 1000 feet of sandstone, shale, and coal in Spring Canyon, it thins eastward by intertonguing with the upper part of the Mancos and disappears as the feather edge of a littoral marine sandstone a few miles southwest of Cottonwood Canyon.

Spieker and Reeside (1925, p. 444) placed the lower boundary of the Blackhawk at the base of the lowest coal bed exposed in the Book Cliffs and the Wasatch Plateau. However the base of the Blackhawk as here redefined is at the base of the Spring Canyon sandstone of Clark. In most outcrops the Blackhawk is overlain disconformably by the Castlegate member of the Price River formation, as seen rather easily at the western end of the Book Cliffs, but only inferred in many other areas where the contact is poorly exposed.

Spieker (1931, p. 36) lists many plant fossils identified by Knowlton, which indicate that the Blackhawk, like the Star Point, is medial Montana (Campanian).

The redefined Blackhawk formation consists of six prominent littoral marine sandstone tongues and many lesser ones, all projecting eastward into the Mancos, where they lose their identity by grading into shale. Above each of them and below the next succeeding littoral marine sandstone, lagoonal deposits of sandstone, shale, and coal were developed behind barrier bars, and where these deposits occur the underlying sandstone is almost everywhere white-capped. This is the case also with the bar sands which appear as sandstone tongues projecting into the Mancos shale.

The six principal members are assigned the following geographic sequence from the highest to the lowest:

Desert member
Grassy member
Sunnyside member
Kenilworth member
Aberdeen member
Spring Canyon member

This division is possible only where the basal littoral marine sandstones are developed. At the extreme western end of the Book Cliffs at Storrs, Utah, only the basal sandstones of of the Spring Canyon and Aberdeen members are present, and between them is about 60 feet of coal-bearing rocks of the Spring Canyon member. Above the basal sandstone of the Aberdeen member is about 800 feet of undifferentiated coal measures of the Blackhawk, which are largely lagoonal but may include some inland flood-plain deposits in the upper portion.

About a mile east of Storrs the basal sandstone of the Kenilworth appears, and the Spring Canyon and Aberdeen members are distinguishable although about 700 feet of undifferentiated coal measures lies above them. The basal sandstone tongue of the Sunnyside member appears near Coal Creek Canyon, so that at that point the Aberdeen and Kenilworth members can be separated from the undivided Blackhawk. Since the basal sandstone tongue of the Grassy member begins at Sunnyside, the Aberdeen, Kenilworth, and Sunnyside members can be identified there while the appearance of the basal sandstone of the Desert member near Desert completes the subdivision of the Blackhawk coal-bearing rocks. Detailed work would probably extend these members westward into the presently undifferentiated beds.

SPRING CANYON MEMBER: The name Spring Canyon coal group was applied by Clark (1928, p. 125) to that portion of the coal-bearing rocks above the Spring Canyon sandstone tongue and below the Aberdeen sandstone. Since this unit is genetically part of the Spring Canyon sandstone cycle, the two units are here

combined to form the Spring Canyon member of the Blackhawk formation.

This member consists of 60–100 feet of coal-bearing shales and sandstones of fresh- and brackish-water origin resting directly upon the massive, basal sandstone of the Spring Canyon, which has an average thickness of about 150 feet in Spring Canyon. Its upper limit in the Book Cliffs is marked by a slight disconformity at the base of a tongue of Mancos shale which separates it from the massive sandstone of the Aberdeen member. This sandstone is absent to the west and southwest, where the member becomes a part of the undifferentiated coal-bearing rocks of the Blackhawk formation.

Two massive offshore bar sandstones, which appear near Helper, occupy almost the entire interval between the basal sandstone of the Aberdeen and the basal sandstone of the Spring Canyon. The coal-bearing rocks of the Spring Canyon member were deposited behind these bars, which mark the eastern limits of the coal-producing swamps. Only a thin tongue of Mancos shale separates the upper bar sandstone and the massive sandstone of the Aberdeen. These sandstones also thin eastward and disappear into the Mancos.

Clark (1928, p. 125–127) and Spieker (1931, p. 79) show that there are at least three important coal beds in the Spring Canyon member. The lowest and most important is the Hiawatha coal named by Spieker (1931, p. 111) for its typical occurrence at Hiawatha in the Wasatch Plateau. It rests directly on the basal sandstone of the Spring Canyon member, or may be separated from it by a few inches or feet of shale and impure fireclay, and ranges from 8 inches to 7 feet 4 inches thick (Clark, 1928, p. 126). To the southwest in the Wasatch Plateau, Spieker (1931, p. 111) reports thicknesses up to 20 feet of good coal. The acid water of the coal-forming swamp which produced this coal formed the white cap of the basal Spring Canyon sandstone.

The lagoonal sides of both offshore bar sandstones are also white-capped and bear the feather edges of coal beds. The major coal swamps evidently formed when the lagoons were almost filled with sediment so that brackish waters were excluded and fresh-water plants flourished.

ABERDEEN MEMBER: The name Aberdeen was first used by Clark (1928, p. 16) in describing the exposures of massive sandstone near the Aberdeen mine north of Kenilworth, Utah, and is here extended to include the overlying coal-bearing rocks and associated offshore bar deposits. It consists of a basal white-capped sandstone with a maximum thickness of 88 feet, and an overlying series of shale, sandstone, and coal with a maximum thickness of about 100 feet at Kenilworth.

The massive, basal sandstone, which disappears at Dugout Creek, is a medium-grained, buff-colored cliff-former, grading downward and eastward into finer-grained, thin-bedded sandstone and silt, and finally into Mancos shale. Foreshore cross-bedding is common in the more massive beds, and, where it is overlain by coal-bearing rocks, an identifying white cap 15–20 feet thick is usually present.

The overlying coal-bearing rocks consist of gray to black shale, lenses of argillaceous, medium- to fine-grained sandstone, and four important coal beds. The lower coal bed, the Aberdeen or Castlegate "A", which rests either directly on the Aberdeen sandstone or is separated from it by a few inches or feet of shale, has been traced westward from Coal Creek Canyon for 20 miles and reaches a maximum thickness of 19 feet at Kenilworth. The Castlegate "B", the Royal Blue, and the Castlegate "C" beds, which occur above the Castlegate "A" in that order, probably are not directly associated with offshore bars.

From Kenilworth to Coal Creek Canyon the coal measures give way to five bar sandstones (Pl. 2, fig. 1) which in turn grade eastward into the Mancos. The second sandstone from the bottom is the most persistent, disappearing at Whitmore Canyon. The bar sandstones are white-capped on the lagoonal side and contain dune, backshore, and lower foreshore laminae.

KENILWORTH MEMBER: The Kenilworth member of the Blackhawk formation here named from exposures near Kenilworth, Utah, consists of a massive, cliff-forming, basal sandstone, an overlying series of coal-bearing rocks, and the offshore bar sandstones behind which they were deposited. The basal sandstone grades into the underlying tongue of Mancos shale,

which rests disconformably on the Aberdeen coal-bearing rocks. It has a sharply defined top, is white-capped where leached by swamp waters, and is medium-grained, thick-bedded, and buff-colored with a maximum thickness of 85 feet. It disappears into the Mancos shale near Sunnyside, although sandstone lenses in the Mancos shale between Woodside and Green River may represent the same horizon. Cross-bedding is common in the more massive beds.

The coal-bearing rocks of the Kenilworth member consist of about 160 feet of gray to black shale with lenses of white and buff sandstone and four minable coal beds, the thickest being the Kenilworth or Castlegate "D" bed which is 19 feet thick at Kenilworth. Above this bed are the Gilson, Fish Creek, and Rock Canyon coals in that order. These coals, too, appear to be developed at horizons corresponding to the top of bar sandstones in the Sunnyside region. (*See* Clark, 1928, p. 128.) These bar sandstones are white-capped where overlain by coal-bearing rocks and are irregularly developed. The two lower bar sandstones formed considerably inland from the seaward edge of the swamp which produced the Kenilworth coal, whereas the upper three formed farther east (Pl. 3). The most eastward extension of the lagoonal rocks is a thin bed of shale and coal which lies on the upper bar sandstone, and which extends almost to Green River. The two lower bar sandstones disappear into the Mancos near Horse Canyon south of Sunnyside, but the upper three continue eastward and split into several tongues, the most extensive of which disappears at Crescent Canyon; a sixth, much less extensive bar sandstone appears near Pace Canyon at the top of the Kenilworth member but disappears near Whitmore Canyon.

SUNNYSIDE MEMBER: The Sunnyside member of the Blackhawk formation, here named for exposures near the town of Sunnyside, Utah, is composed of a massive, basal sandstone tongue and the overlying coal-bearing rocks which are replaced eastward by offshore bar sandstones.

Like the other littoral marine sandstones of the Blackhawk this basal sandstone is medium-grained, buff, and massive. It reaches a maximum thickness of about 50 feet at Pace Canyon, grading downward and eastward into a shale tongue, which in turn rests disconformably on the coal-bearing rocks of the Kenilworth member. Where overlain by coal-bearing rocks the basal Sunnyside sandstone, too, is white-capped. It first appears at Kenilworth and splits eastward into two tongues which disappear into the Mancos shale near Horse Canyon south of Sunnyside.

The coal-bearing rocks of the Sunnyside member are comparatively thin, consisting of about 25 feet of sandstone, shale, and coal which, at a point about 4 miles northeast of Desert, are replaced by the last of a series of six offshore bar sandstones beginning near Sunnyside. The highest and most extensive of these sandstones disappears into the Mancos near Horse Canyon east of Green River, while the lowermost bar sandstone splits into two tongues and disappears about 10 miles south of Sunnyside. About 6 miles farther south along the cliffs, two stringers of sandstone appear near the horizon of the upper tongue and continue east to Crescent Canyon. The gap in outcrop is probably due to an embayment in the old shore line.

Two important coal beds are present in the Sunnyside member, the Upper and Lower Sunnyside coals. The lower one rests directly on the white cap of the basal sandstone tongue near Sunnyside, but from here to Horse Canyon it is overlapped by the first bar sandstone and is relatively thin. The maximum thickness of this coal is 5 feet.

The Upper Sunnyside coal, which is mined at Sunnyside, lies 30–35 feet above the lower coal bed and can be traced southward to the Price River. It is about 22 feet thick at Horse Canyon where it is split near the top. It is a composite of several coal beds and splits southward into 3 subsidiary beds (Pl. 3) where offshore bars appear. In each case the lower portion splits off and is overlapped by a bar sandstone. The splitting is indicative of the mode of formation. The first of the thinner coals formed in a shallow lagoon and extended inland for a considerable distance. Then a sharp pulse of subsidence of small magnitude caused the barrier bar to be submerged and allowed the marine waters to inundate that portion of the lagoon nearest the bar. A new

barrier bar must have started to form immediately, and soon the vegetation, which had continued to flourish inland, again invaded the outer parts of the lagoonal area.

This process of inundation and swamp regeneration often resulted in the formation of a thick coal bed in that portion of the swamp near the shore which was little affected by the inundations. Such a coal bed gradually thins landward and splits seaward into several thinner beds separated by thin lagoonal deposits and thin inner edges of offshore bar sandstone. The most striking example of this process is that of the Upper Sunnyside coal; however, the same process operated during Aberdeen and Kenilworth time. (Pl. 3).

GRASSY MEMBER: The rocks here designated the Grassy member of the Blackhawk formation receive their name from exposures in the vertical cliffs east of the railroad siding of Grassy. This member consists of a basal littoral marine sandstone with a maximum thickness of about 60 feet and a series of coal-bearing rocks with a maximum thickness of about 50 feet.

The basal white-capped sandstone of this member first appears at Sunnyside and fades into the Mancos shale at Coal Canyon. Its lithology is similar to that of the other littoral marine sandstones of the Blackhawk. It is a medium-grained, buff sandstone, which grades downward and eastward from massive into thin-bedded sandstone and finally into a marine shale tongue (Pl. 1, fig. 1). This shale rests disconformably on coal-bearing rocks of the Sunnyside member.

No minable coals are found in the Grassy member though some thin beds are present. Coal-bearing rocks of this member are absent east of Coal Canyon.

Between Green River and Coal Canyon three offshore bar sandstones appear and form the eastern limit of the coal-bearing unit. The most persistent is the middle sandstone, which disappears at Crescent Canyon.

DESERT MEMBER: The uppermost of the more extensive sandstone tongues appears near Desert. This sandstone and the overlying coal-bearing rocks which separate it from the Castlegate member of the Price River, are here defined as the Desert member of the Blackhawk, named for exposures in the nearly vertical cliffs of Mt. Elliot east of Desert, a siding on the D.&R.G.W.R.R.

The basal unit of this member is a medium-grained, buff sandstone with a maximum thickness of about 70 feet. It is massive near the top, but grades downward into thinner beds and finally into Mancos shale which in turn rests disconformably on the coal measures of the Grassy member. Eastward it disappears into Mancos at Saleratus Canyon.

Many thin, unminable lenses of coal are present, but their form and the absence of a well-defined white-cap on the basal sandstone seem to indicate the lack of any large coal swamp at the time of their deposition.

About 50 feet of coal-bearing rocks is present at Desert, but it thins rapidly to the southeast and is absent at Green River where the basal sandstone of the Castlegate rests directly on the basal sandstone of the Desert member. This was probably a local high during the deposition or an area of post-Blackhawk erosion since the coal measures reappear near Tuscher Canyon as a thin wedge of sandstone, shale, and coal thickening to the east. These rocks reach a thickness of about 80 feet near Horse Canyon east of Green River where they abut against the lagoonal side of the first two of a group of nine offshore bar sandstones, which terminate the strata just east of Thompson Canyon. An excellent cross section of a lagoon on a small scale is exposed in the cliffs between Coal Canyon and Thompson Canyon and another directly above these deposits in the Castlegate member of the Price River.

The uppermost of the nine bar sandstones disappears into the Mancos shale a short distance southwest of Cottonwood Canyon.

PRICE RIVER FORMATION

The name Price River formation was applied by Spieker and Reeside (1925, p. 445) to a series of noncoal-bearing beds above the Blackhawk formation. It includes the upper part of the Mesaverde group of Clark (1928, p. 20), and derives its name from exposures in Price River Canyon west of Castlegate. In the western portion of the Book Cliffs it consists of a massive basal sandstone (the Castlegate member) and overlying shale, sandy shale, and

lenslike beds of massive, coarse- to medium-grained sandstones.

Farther east, in the Beckwith Plateau, the basal Castlegate member is separated from the remainder of the Price River by a thin wedge of Mancos shale (Buck tongue), which thickens eastward. The Castlegate thins eastward and disappears in the Mancos near West Salt Creek. Simultaneous with the deposition of the Buck tongue the thick basal sandstone of the Sego member was deposited above the shale. This member includes several sandstone tongues and related coal-bearing rocks.

In the vicinity of Thompsons, D. J. Fisher (1936, p. 19) subdivided that part of the Price River above the Sego sandstone tongues into two members: the lower, coal-bearing member, called the Neslen for exposures in Neslen Canyon, and the upper, noncoal-bearing member, called the Farrer for exposures near the Farrer Mine in Coal Canyon. Erdmann (1934, p. 32) in western Colorado subdivided the Price River equivalents above the Sego member into two units: the lower, called the Mt. Garfield formation for exposures on Mt. Garfield near Palisade, included the coal-bearing rocks and the lower part of the noncoal-bearing rocks and correlated with the Bowie and Paonia shale members of the Mesaverde to the south of Grand Mesa; the remainder of the noncoal-bearing rocks, called the Hunter Canyon formation for exposures in Hunter Canyon, correlated with the undifferentiated Mesaverde of Lee.

The distinction between the Mt. Garfield and Hunter Canyon formations is of the same order as that between the Neslen and Farrer members, and, since there is obviously no need for both sets of terms, the writer proposes to drop the terms Mt. Garfield and Hunter Canyon and to adopt the term Neslen facies for all the coal-bearing rocks of the Price River, and the term Farrer facies for the noncoal-bearing rocks above the Neslen facies. Measurements by Fisher and Erdmann show that the passage between coal-bearing and noncoal-bearing strata is gradational and does not occur at one widespread stratigraphic horizon. Since coal-bearing and noncoal-bearing rocks as well as littoral marine sandstones are characteristic of related environments, shifting of one environment would result in a shifting of the other. The boundary between the Neslen and Farrer facies should be nearly as zigzag as that between the littoral marine sandstones and marine shales. Thus it can be seen that this boundary rises stratigraphically to the east (Fig. 2). The Neslen facies remains fairly uniform in thickness despite intertonguing to the east with littoral marine sandstones and marine shales, whereas the rocks of the Farrer facies thicken eastward.

Three prominent littoral marine sandstones appear in the rocks of the Neslen facies above the Sego member in western Colorado. The lower of these with its associated lagoonal deposits is here named the Corcoran member, the middle member is named the Cozzette member, and the upper is named the Cameo member, replacing the term Rollins sandstone member of Lee (1912, p. 19).

At the extreme western end of the Book Cliffs the Price River formation rests disconformably on the Blackhawk formation more obviously than in other regions. The contact with the overlying North Horn is gradational in Price River Canyon and is drawn where red beds first appear in the geologic section. East of the Green River the Price River formation is overlain unconformably by the Tuscher formation, which is probably basal North Horn.

The Price River formation ranges in thickness from about 1000 to 2000 feet at the western end of the Book Cliffs and from 1000 to 2500 feet in western Colorado. Fossil evidence reviewed by Spieker (1946, p. 131) indicates Maestrichtian stage. However, Cobban and Reeside (1952) place it somewhat lower in the Campanian and Maestrichtian stages.

Neslen Facies

Neslen facies is the name applied to the coal-bearing rocks of the Price River formation, and is taken from exposures in Neslen Canyon, near Thompsons, Utah. D. J. Fisher (1936, p. 16) first used the name Neslen as a stratigraphic term, but, since this is a facies development which laterally crosses time lines, the use of a formational name would be inconsistent with the writer's delimitation of more or less contemporaneous members. The name is retained but is used to identify a facies.

By definition the rocks of the Neslen facies are coal-bearing though coal makes up only a small portion of the rocks, which are predominantly shale and sandstone. The shales are largely nonmarine and are gray to black, depending on their carbon content. The sandstones are of two types, the most common are the thin, lenticular, buff sandstones found in the lagoonal deposits, and the most conspicuous are the massive, buff, littoral marine sandstones which are traceable for many miles. These sandstones, like those of the Blackhawk, are used to subdivide the rocks of the Neslen facies into five members.

Erosion is indicated by slight disconformities at the base of the Mancos shale tongues which underlie each of the littoral marine sandstones and which formed simultaneously with the overlying sandstones.

The Castlegate member belongs to both the Neslen and Farrer facies. However, since it does include coal-bearing rocks throughout a considerable portion of its outcrop, it may be discussed as a member of the Neslen facies.

The five members of the Neslen facies are, in normal stratigraphic sequence from top to bottom:

Cameo member
Cozzette member
Corcoran member
Sego member
Castlegate member

Each member consists of one or more basal sandstone tongues and associated lagoonal and barrier bar deposits and can be distinguished only where the littoral marine sandstones are present. Thus the landward, lagoonal deposits of the Neslen facies, not penetrated by littoral marine sandstone, have not been subdivided. Rocks of the Neslen facies range in thickness from about 100 feet near Woodside to about 1000 feet near the Colorado-Utah line.

CASTLEGATE MEMBER: Clark (1928, p. 119) applied the name Castlegate sandstone member to the massive, basal sandstone of the Price River formation, for exposures near the town of Castlegate, Utah. Since its name does not convey the idea that this unit also contains some coal-bearing rocks, it is here called the Castlegate member of the Price River. It consists primarily of a massive, cross-bedded, white or pink sandstone which changes from coarse- to fine-grained as traced eastward. This portion of the Castlegate, belonging to the Farrer facies, represents an inland flood-plain deposit, which is replaced by lagoonal deposits for a few miles near Horse Canyon east of Green River. These lagoonal deposits, which contain only a few thin coals, are replaced eastward in the vicinity of Thompsons by littoral marine, bar sandstones. This portion of the Castlegate, as well as a thin topmost lagoonal deposit present from Desert east to Cottonwood Canyon, is placed in the Neslen facies.

The near absence of coal in the Castlegate seems to indicate that, as Blackhawk time proceeded, coal-forming conditions waned and reached a low in the Castlegate only to resume again in the Neslen.

From the western end of the Book Cliffs to Green River, the Castlegate member is composed mainly of nonmarine sandstone which formed in response to an orogenic pulse (Spieker, 1946, p. 159). At the eastward edge, at Horse Canyon near Green River, it passes into lagoonal deposits. Three littoral marine bar sandstones, form the eastern margin of the lagoonal deposits and interfinger eastward with the Mancos. The upper two bars are the most persistent but disappear at West Salt Creek.

Microscopic examinations show that the sand of the inland flood-plain portion of the Castlegate (Farrer facies) derives its white color both from kaolinized feldspars and a calcareous cement. The littoral marine sandstones are fine- to medium-grained and are buff in color. Cross-bedding in the inland flood-plain portion is of the torrential type, while that of the littoral marine portion is dominantly marine.

This member ranges from about 500 feet thick at Castlegate to a feather edge near West Salt Creek. The coal-bearing rocks, which reach a maximum thickness of about 90 feet near Crescent Canyon, contain no workable coals.

The disconformity noted at the western end

PLATE 1.—UPPER CRETACEOUS, BOOK CLIFFS, UTAH-COLORADO

FIGURE 1.—BASAL SANDSTONE OF THE GRASSY MEMBER NEAR GREENRIVER, UTAH

FIGURE 2.—DISCONFORMITY AT BASE OF CORCORAN MEMBER NEAR FARMERS MINE

FIGURE 2

FIGURE 1

UPPER CRETACEOUS, BOOK CLIFFS, UTAH–COLORADO

FIGURE 1

FIGURE 2

FIGURE 3

UPPER CRETACEOUS, BOOK CLIFFS, UTAH–COLORADO

of the Book Cliffs east of Kenilworth and west of Thompson Canyon may be present, but has not as yet been definitely recognized. Some reworked Blackhawk material may have been mixed with typical Castlegate and redeposited as a transitional zone in some places. This seems to be the case near Sunnyside, although inaccessible exposures prevented an accurate investigation. In Thompson Canyon the lower bar sandstone grades downward into a thin shale tongue which rests disconformably on the underlying coal-bearing rocks of the Blackhawk. In Coal Canyon the basal contact appears to be disconformable, and in the canyons between Green River and Crescent Junction numerous springs are present at the contact and suggest an unconformity.

The contact with the overlying sandstones and shales of the Farrer facies in the western Book Cliffs is poorly defined but is usually placed at the top of the massive, basal sandstone of the Price River formation. Since the top is transitional west of Woodside, it is not everywhere the same age. Where the Buck tongue overlies the Castlegate member east of Woodside, the Mancos shale rests in most places, apparently disconformably, on coal-bearing rocks. The disconformity noted by D. J. Fisher (1936, p. 15) in Horse Canyon east of Green River is probably at the contact between the coal-bearing rocks and the underlying basal sandstone of the Castlegate.

Fossils from the Castlegate member (Spieker and Reeside, 1925, p. 446; D. J. Fisher, 1936, p. 14–15) indicate a late Montana age.

SEGO MEMBER: Sego member is the term here introduced to replace the name Sego sandstone member of the Price River formation applied by D. J. Fisher (1936, p. 15) to a series of interbedded sandstones and shales at the coal mining town of Sego near Thompsons, Utah. It includes the rocks formerly called the Sego sandstone as well as the overlying and closely related coal-bearing rocks. Its base is drawn at the gradational contact of the lowermost littoral marine sandstone tongue with the underlying marine shale. The upper boundary is placed at the slight disconformity between the coal-bearing rocks and the overlying marine shale tongue beneath the Corcoran member. The average thickness is 200 feet. The many fossils in this unit indicate a Campanian age.

The littoral marine portion consists of several thin sandstone tongues; the first appears near Woodside. Three are more persistent than the rest and can be traced eastward into Colorado, and of these three the upper is the most prominent. Both the middle and lower tongues disappear into the Mancos north of Nash Canyon but reappear at Cottonwood Canyon; probably because of a local embayment or indentation of the old shore line, since the outcrops in this part of the cliffs trend northeast approximately parallel to the old shore line. The crossbedding of the Sego sandstone in this area indicates a northwestern source. The middle sandstone tongue pinches out twice more before finally disappearing at Camp Canyon, while the lower tongue continues eastward from Cottonwood Canyon and finally disappears at Big Salt Wash, and the upper tongue is continuous from Saleratus Canyon to a point just east of the Anchor No. 1 Mine.

In most localities each main sandstone tongue is overlain by fresh- or brackish-water deposits of sandstone, sandy shale, or shale, and where this is a massive channel sandstone, the actual thickness of the littoral marine sandstone is difficult to ascertain. Thin westward-pointing tongues of Mancos shale grading upward into the overlying littoral marine sandstone tongue rest disconformably on the brackish- or fresh-water deposits. Those deposits above the lower tongue are formed behind a series of poorly defined bar sandstones between West Salt Creek and Camp Canyon. In the embayment near Cottonwood Canyon, no trace of barrier bars was noted. The nonmarine deposits above

PLATE 2.—UPPER CRETACEOUS, BOOK CLIFFS, UTAH-COLORADO

FIGURE 1.—OFFSHORE BARS OF ABERDEEN MEMBER IN COAL CREEK CANYON

White-capped, basal sandstone of Kenilworth member (k) rests almost directly on uppermost of five offshore bar sandstones of Aberdeen member (a_1, a_2, a_3, a_4, a_5)

FIGURE 2.—SEGO MEMBER IN EAST SALT CREEK

Upper and lower tongues of basal sandstone are separated by Anchor Mine shale tongue

FIGURE 3.—SEGO MEMBER AND ANCHOR COAL IN BIG SALT WASH

Stringers of lower tongue remain near middle of shale slope. Upper tongue forms steep cliff under coal. Basal Corcoran sandstone immediately overlies coal

the middle tongue are spotty, but those above the upper tongue are continuous and important. Where the upper part of this member is delimited by the overlying marine shale tongue and littoral marine sandstone, these coal-bearing rocks have an average thickness of 60 feet. The coal-bearing rocks above the upper sandstone tongue formed behind barrier bars which appear in the cliffs between the Farmers Mine and the Book Cliffs Mine.

The littoral marine sandstones vary from massive-bedded, medium-grained, buff sandstones to thin-bedded, fine-grained, silty gray sandstones with shale partings. The predominant quartz particles are angular, and many show pitting. The maximum thickness of the sandstones is about 50 feet; however, channel sandstones may greatly increase the apparent thickness. These brackish- or fresh-water sandstones are lighter in color than the littoral marine sandstones and give the false appearance of white capping as is typical of littoral marine sandstone under coal. Cross-bedding of the channel sandstones is fluviatile in most places; and the sand grains composing this upper portion are larger, fresher, and more angular than those of the littoral marine sandstone. Pitting of grain surfaces is common to both, however, while in some places these lenslike sandstones exhibit eolian cross-bedding, and the grains are somewhat frosted. The presence of flakes of muscovite seems to preclude any great amount of wind action. In several localities channeling effects were noted as at one exposure in Neslen Canyon where a channel cuts through the littoral marine sandstone and into the Mancos below.

The 60 feet of coal-bearing rocks above the upper sandstone tongue of the Sego member are of most concern, since no coal beds of importance are found in the irregular, fresh- and brackish-water deposits below the upper tongue. The coal-bearing rocks consist of sandstone, sandy shale, carbonaceous shale, and one important coal bed. Erdmann and D. J. Fisher indicated that this coal bed could be traced from near Green River to a point near Hunter Canyon; however, because of miscorrelation by Erdmann at Big Salt Wash, a confused terminology has developed.

Erdmann (1934, p. 78) gave the name Palisade coal to a coal mined near Palisade. He traced this coal westward to the Utah line to where Fisher had traced it from the Green River. Since it had been traced along its entire outcrop, the deposit was believed to be continuous. The nomenclature that was adopted and the resulting wide extent of the Palisade coal are not justified because the coal Fisher traced to the Colorado line as the Palisade coal is actually the same as Erdmann's Anchor coal.

In the Anchor Mine tongue at Big Salt Wash Erdmann (1934, p. 36) reports a series of coal-bearing rocks above the lower sandstone tongue of the Sego. The most conspicuous unit is a coal bed which Erdmann named the Anchor coal. Investigations by the writer show no such coal-bearing unit in the Anchor Mine tongue. Actually, these beds lie above the upper sandstone tongue of the Sego and continue eastward to the vicinity of Hunter Canyon where they are replaced by barrier bars. The Palisade coal bed is about 100 feet above the Anchor coal bed and can be traced only a short distance west of Big Salt Wash; it may correlate with the Ballard coal zone of Fisher.

This confusion arises because the lower sandstone of the Sego thins rapidly from East Salt Creek to Big Salt Wash (Pl. 2, figs. 2, 3).

The writer proposes to use Erdmann's name Anchor coal for the main coal bed of the Sego member, named for the Anchor Mine where it is about 5 feet thick. As redefined, it corresponds to the Palisade coal bed of Fisher and the Anchor coal bed of Erdmann. Lithologically the offshore bar sandstones which terminate the coal measures are similar to those of the Blackhawk, but white caps are rare.

CORCORAN MEMBER: The name Corcoran member is here applied to the unit of littoral marine sandstone and associated coal-bearing rocks which lies above the Sego member, and is separated from it by a thin tongue of Mancos shale with a disconformity at its base (Pl. 1, fig. 2). It takes its name from exposures near the old Corcoran mine north of Palisade, Colorado, and has an average thickness of about 100 feet.

At the base of this unit a massive littoral marine sandstone tongue grades downward and eastward into the Mancos and is lithologically

similar to those of the Blackhawk and of the Sego; it first appears near Big Salt Wash and is traceable eastward into Grand Mesa, where it disappears. Where overlain by coal measures it is commonly pure white and miners call it the "White Pioneer". A series of offshore bar sandstones appears above it near Watson Creek in Grand Mesa and forms the eastern limit of the 50 feet or so of coal-bearing rocks.

Erdmann (1934, p. 81) describes the main coal zone of this member as a series of overlapping and isolated lenses of coal in a thick deposit of carbonaceous shale. This is the Palisade coal of Erdmann, a term here restricted to the coal of the Corcoran member, which is traceable from the vicinity of Big Salt Wash to Plateau Creek. Four separate beds can be traced eastward from Hunter Canyon to near Plateau Creek where they pass under the Colorado River and continue under Grand Mesa. Each is 1–4 feet thick and each presumably corresponds to the top of an offshore bar sandstone in the vicinity of Watson Creek.

COZZETTE MEMBER: The littoral marine sandstone and associated coal-bearing rocks which overlie the Corcoran member are here named the Cozzette member of the Price River formation for exposures near the old Cozzette mine north of Palisade. Its basal sandstone grades downward into a tongue of Mancos, which in turn rests disconformably on the coal-bearing rocks of the Corcoran member. The upper limit is drawn at the disconformity between the coal-bearing rocks and the overlying tongue of Mancos shale.

In the outcrops observed, the basal sandstone appears to be composed of two units with a combined thickness of about 130 feet. This massive-bedded, buff sandstone begins near Hunter Canyon and extends into Grand Mesa. The extent of the sandstone and the position of the associated offshore bar sandstones remain to be accurately determined.

CAMEO MEMBER: The uppermost coal-bearing rocks comprise the Cameo member of the Price River formation, named for exposures near the town of Cameo. This member is limited below by a basal littoral marine sandstone which at one time was believed to be the Rollins sandstone of Lee. However, tracing by W. W. Boyer (Spieker, 1949, p. 56) indicated that the true Rollins is much higher stratigraphically. The upper limit of the Cameo member is the boundary between the coal-bearing rocks of this member and the noncoal-bearing rocks of the Farrer facies. This upper boundary is, therefore, indefinite but will serve until the remainder of the intertonguing details are determined. There are undoubtedly other littoral marine tongues in the coal-bearing rocks above the Cameo sandstone tongue in Grand Mesa which will facilitate further subdivision of these coal-bearing rocks.

The massive, white-capped, basal sandstone of the Cameo member is conspicuous in the cliffs near Palisade and in Grand Mesa. It is this thick-bedded, medium-grained cliff-forming sandstone which caps Mt. Garfield near Palisade, first appearing near Hunter Canyon and continuing south beyond Grand Mesa. It grades downward into a shale tongue which lies disconformably on the coal measures of the Cozzette member.

Near Mt. Garfield about 250 feet of coal-bearing rocks consisting of sandstone, sandy shale, carbonaceous shale, and coal lies above this massive sandstone. Resting directly on the basal sandstone is the Cameo coal zone, the most important coal zone in the eastern Book Cliffs. Erdmann (1934, p. 83) states that the Chesterfield coal of Utah is probably equivalent to the Cameo coal. In most exposures the Cameo zone consists of two beds separated by a thick parting. The upper bed is less bony and approaches 20 feet in thickness (Erdmann, 1934, p. 84).

About 60 feet above the base of the Cameo coal zone is the Carbonera coal zone (Erdmann, 1934, p. 86), a series of discontinuous lenses of 10 feet maximum thickness.

Offshore bar sandstones undoubtedly exist farther to the east and southeast, but none is present in the region studied.

Farrer Facies

The name Farrer formation was applied by D. J. Fisher (1936, p. 19) to the noncoal-bearing rocks of the Price River formation in Coal Canyon and is here retained as a facies term applied to the noncoal-bearing rocks of the Price River.

These beds of coarse to fine sandstone, shale, and sandy shale do not contain appreciable coal although a few thin beds are found at various horizons. The lenslike sandstones of this facies, which constitute two-thirds to three-fourths of the bulk of the rocks, are composed largely of subangular quartz grains and are cemented by mud, calcium carbonate, and ferruginous cements. As a result, they are somewhat darker than the sandstones in the Neslen facies. Most of the sandstones in the western portions of the Book Cliffs are fairly light in color, and some are even white because of the kaolinization of feldspars and leaching of iron. Shales of this facies are mostly gray, but in places they are olive green.

West of Woodside, the Castlegate member, as well as the remainder of the Price River formation, is noncoal-bearing and thus should be placed in the Farrer facies. East of Woodside, however, the coal-bearing rocks begin to appear at stratigraphically higher levels. Therefore, the contact between the two facies must rise in a zigzag manner toward the east. This rise should normally be expected to decrease the thickness of the rocks of the Farrer facies, but the Farrer rocks thicken from about 1200 feet in Price River Canyon to about 1500 feet near Palisade (Spieker, unpublished data). The thickening must be due largely to greater deposition to the east, since the 1200 feet in Price River Canyon comprises the whole of the Price River formation, while the 1500 feet near Palisade is only the upper part. The increase for the entire Price River is much greater since the interval between the Castlegate horizon and the North Horn equivalents is about 3000 feet near Palisade.

It follows then that west of Woodside the base of the Farrer facies is the base of the Castlegate member. East of that point it is a rather indefinite line marking the transition from coal-bearing to noncoal-bearing rocks. The upper boundary coincides with that of the Price River. In Price River Canyon the passage to the North Horn formation is transitional, and the boundary is arbitrarily put at the base of the lowermost red shale bed in the section. East of Green River the Price River is disconformably overlain by a unit which Fisher (1936, p. 20) named the Tuscher. In Colorado Erdmann (1934, p. 53) noted a disconformity between the top of the Price River (his Hunter Canyon) and the overlying Ohio Creek Conglomerate.

No subdivision of the Farrer was attempted beyond the separation of the Castlegate member section.

North Horn Formation

The North Horn formation was named by Spieker (1946, p. 132) for exposures on North Horn Mountain in the Wasatch Plateau. Rocks of this unit had previously been treated as the lower member of the Wasatch formation. Consisting of variegated shale and sandstone, conglomerate, and fresh-water limestone which represent fluviatile and lacustrine conditions (Spieker, 1946, p. 133), it contains a few thin beds of poor coal in Price Canyon where the formation is about 2200 feet thick.

In the western portion of the Book Cliffs the North Horn grades downward into the Price River conglomerate and sandstone and passes transitionally upward into the Flagstaff limestone of Paleocene and Eocene age (Spieker, 1946, p. 133). In the vicinity of Green River, Spieker (1946, p. 142) found that the basal part of the North Horn is probably equivalent to the Tuscher formation of Fisher. The Tuscher is a light-colored, conglomeratic sandstone about 200 feet thick. It rests disconformably on noncoal-bearing rocks of the Price River formation, but this disconformity has not been recognized farther west. In the canyon of Green River a strong disconformity separates the North Horn from the overlying Colton red beds (Spieker, 1946, p. 142). Traced eastward the Tuscher of Fischer is probably continuous with the Ohio Creek conglomerate of Erdmann which may be the same as the Ohio Creek of Lee (Erdmann, 1934, chart opp. p. 26). Both are lithologically the same and both rest disconformably on the Price River. At the eastern end of the Book Cliffs the Ohio Creek is overlain disconformably by the Wasatch (Ruby) formation which is probably equivalent to the Colton of central Utah.

The part of the column embraced by the North Horn thins from 2200 feet near Castlegate to about 200 feet in Grand Mesa at the

eastern end of the Book Cliffs. Fossil evidence (Spieker, 1946, p. 134) shows the North Horn to include beds of both Lance and Paleocene age. The lower part has yielded dinosaurian remains, the middle part has yielded no diagnostic vertebrate fossils, but the upper part contains undoubted Paleocene mammalian remains. Therefore, the Mesozoic-Cenozoic boundary must lie within the middle part of the North Horn.

Facies

Spieker (1949, p. 60) recognizes five environments in the Upper Cretaceous rocks of central and eastern Utah, of which one (piedmont) is not present in the Book Cliffs. The other four are as follows:

Sedimentary facies	*Environments*
(1) a. Conglomeratic sandstone b. Variegated beds, clay shale, fresh-water limestone, sandstone, and some coal c. Buff sandstone and gray shale	(Inland flood plain, channel and lake).
(2) Coal-bearing successions of buff to gray sandstone and shale	(Lagoonal, estuarian, flood plain, and swamp)
(3) Massive and evenly bedded medium to fine buff sandstone (partially white-capped)	(Littoral marine)
(4) Gray shale and siltstone, evenly bedded	(Offshore marine)

Each of these four environments is characterized by certain types of sediments which constitute a sedimentary facies, but since none of these lithofacies consists of a single lithologic unit, the writer follows Spieker (1949, p. 60) in designating them in the environmental terms of inland, lagoonal, littoral marine, and marine. In discussing the Price River formation it is not possible to treat each facies separately. Therefore, the lagoonal and littoral marine facies are combined in the Neslen, and the inland facies designated the Farrer facies.

These four environments, in response to the subsidence of the basin and the regression of the sea, shifted first seaward then landward and resulted in intricate intertonguing of the deposits of the several environments or facies (Fig. 3). Though intertonguing is common to all facies, it is best exemplified in the littoral marine and marine facies. Long tongues of littoral marine sandstone extend eastward into the Mancos shale, and long thin tongues of the Mancos penetrate westward into the coal-bearing rocks and separate the tongues of littoral marine sandstone. Although intertonguing between the inland, lagoonal, and littoral marine facies is more difficult to follow, it is comparable to that of the littoral marine-intertonguing.

The four facies are traceable in a normal succession vertically throughout the Book Cliffs, and laterally from inland on the west to offshore marine on the east. Shifting of environments has caused the succession to be partially repeated in some areas. These shifts and the accompanying shifts of the strand line, though not steady, ultimately resulted in the filling of the basin and the withdrawal of the sea from this region.

The source of these thick deposits was the rising Meso-cordilleran geanticline to the west and northwest. The intertonguing and the cross-bedding in the non-marine deposits point to a northwesterly source. This accords with the fact that these deposits show little variation along northeastward trending exposures which parallel the old shore line.

Intertonguing

General Nature

The boundary between the inland and lagoonal facies is poorly defined, since the two environments may be gradational, and their sediments are almost identical. Consequently, only its approximate position is indicated in Figure 3. The boundary between the lagoonal and littoral marine deposits, and the one between the littoral marine and marine deposits were traced out in detail. (Pl. 3).

The best example of intertonguing of these facies is that between the littoral marine sandstones and the marine shale. Clark (1928, Pl. 4) first observed it in the western Book Cliffs, where tongues of sandstone of the Star Point sandstone and Blackhawk formation disappear

eastward into the Mancos. Later work by Spieker, Baker, Reeside, Fisher, and Erdmann showed continuation eastward to Grand Mesa and beyond, where at least 13 major littoral marine sandstone tongues and many lesser offshore bar sandstone tongues penetrate the shale. The lowest tongue reaches the shortest distance east, while each succeeding one reaches farther, because of the eastward recession of the sea shore.

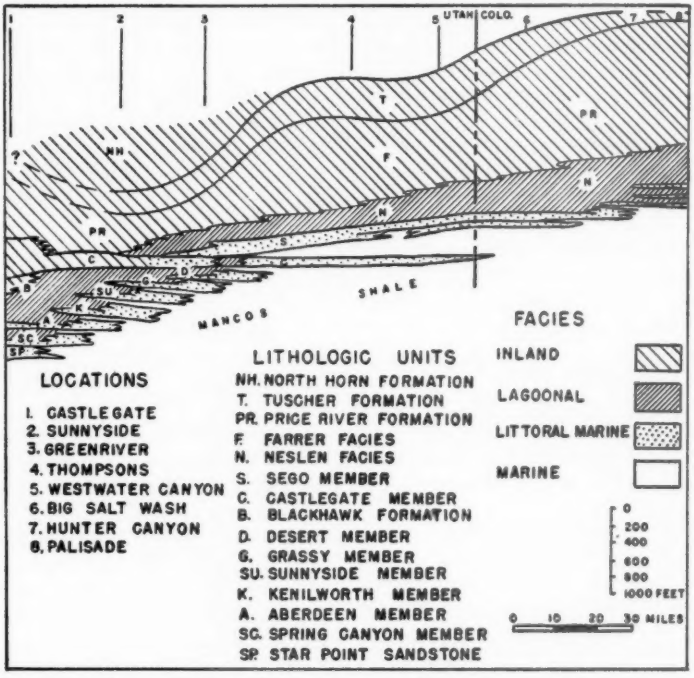

FIGURE 3.—DISTRIBUTION OF FACIES IN UPPER CRETACEOUS OF BOOK CLIFFS
Adapted from Spieker (1949).

Since these sandstones are in reality old beach deposits, tracing their upper surfaces is equivalent to tracing out the changing position of the old beach line. As the beach line shifted the sand-mud contact must also have shifted with each marine transgression resulting in the landward migration of both the shore line and the sand-mud line. As transgression ceased, regression must have begun almost immediately. Regression was probably caused and certainly accompanied by regressive deposits of littoral marine sandstone and marine shale, and at the base of each littoral marine sandstone tongue is a corresponding thin tongue of marine shale which points westward and which extends almost to the initial point of the littoral marine sandstone (Pl. 3). The sequence of environments in coal-forming times from west to east was from inland flood plain to lagoonal to littoral marine to shallow sea. The sand extended out to an unknown depth where it graded into mud, and the line of contact now appears at the base of the sandstone tongues. Shore-sediment changes in the Upper Cretaceous must have been more regular than those along present-day shores, to produce such widespread tongues of sandstone with such uniform thickness.

Marine Shale

The marine shale of the Upper Cretaceous of the Book Cliffs is the Mancos shale. Where deposited far from shore it consists mainly of

thinly laminated, light-gray to gray-black clay shale, with a few thin bentonite beds and an occasional thin limestone stringer.

In near-shore areas the Mancos is more sandy and is lighter in color. Littoral marine sandstone tongues grade laterally and downward into the shale. Since the base of the sandstone is at the line where sandstone becomes dominant over shale, many sandstone stringers are included in the Mancos. When the tip of a sandstone tongue disappears into the shale, a sandy zone distinguishable by its lighter color, continues for a considerable distance into the shale. This sandy zone may bear ironstone concretions, which in some cases contain well-preserved marine fossils. Plant fragments are common in the thin tongues of the Mancos which separate the sandstones from the underlying coal-bearing rocks. Farther east lenses and thin stringers of limestone replace the sandy zone.

Littoral Marine Sandstones

The littoral marine sandstones, the most conspicuous element of the Book Cliffs, vary greatly in thickness but are quite persistent. They consist of medium- to fine-grained, buff, massive or thin-bedded, quartz sandstone. The large amount of heavy minerals in some parts of these sandstones probably resulted from the sorting action of the wind and waves on the upper foreshore and backshore portions of the beaches. A few flakes of muscovite appear in most samples, and pitting of the quartz grains is negligible. The cementing material is primarily calcareous, but the upper portion may contain large amounts of ferruginous cement where the sandstone is overlain by marine shale. Where it is overlain by coal-bearing rocks, a white cap averaging 15 feet thick is present. The leaching action of the swamp waters carried away the ferruginous portion of the cement but left the calcium carbonate portion intact. In some places the iron was carried downward and deposited near the bottom of the leached zone, and this removal has in places weakened the bond between the sand grains and allowed pitted surfaces to develop in the white-cap portion. These sandstones have sharply defined tops in contrast to the gradational bases, and this combination produces the typical steplike profile of the desert (Spieker, 1949, p. 64). The flat tops may be preserved by an increase of ferruginous cement, but they were produced as subaerial gradation planes truncating successive beach planes as the land built seaward. The truncation was by currents transporting sand seaward, as testified by local channeling, and by the action of waves and ocean currents.

The transitional bases of the sandstones represent the position of the sand-mud transition which migrated seaward at an essentially constant level. The diagonal bedding planes represent successive beach and bottom surfaces. They are farthest apart near the top surface of the sandstone, producing massive bedding, and they converge toward the base to form thin-bedded sandstone and then pass from the sandstone into siltstone and shale. This is apparent only where the beaches were sufficiently steep to allow tracing of individual beds, as in the Panther sandstone west of Helper. Seaward (eastward) the bedding planes become more nearly horizontal and closer together. Shale partings appear, and the sandstone grades finally into shale. If each bedding plane be considered a fairly well-defined time line, then the top of each sandstone must represent innumerable time lines which split at regular intervals and cut diagonally through the sandstone tongue. Each line flattens out in the shale, to run parallel to the top of the tongue, but converging slightly toward its fellows seaward because of the decrease of sediment. Such a sandstone bed is seen in Figure 4 which also depicts the mode of formation of an offshore bar.

Spieker (1949, p. 65) has concluded that during the deposition of each sandstone tongue the local basement must have remained stable but that, in order to produce the next higher shale and littoral marine tongues, subsidence must have occurred, and that so rapid as to have prevented the formation of transgressive deposits. Its only effect was a slight disturbance of the poorly consolidated shales and sandstones of the lagoonal facies. Spieker (1949, p. 66) suggested that there is no transgressive littoral marine sandstone in the Book Cliffs and investigations support this conclusion.

Sears *et al.* (1941, p. 104–105), in their study of intertonguing of the Upper Cretaceous deposits of the southern San Juan Basin, report both transgressive and regressive deposits in-

the Book Cliffs are probably regressive deposits formed by the progressive filling of a shallow marine basin. The shale tongues between them must then be caused by sharp pulses of sub-

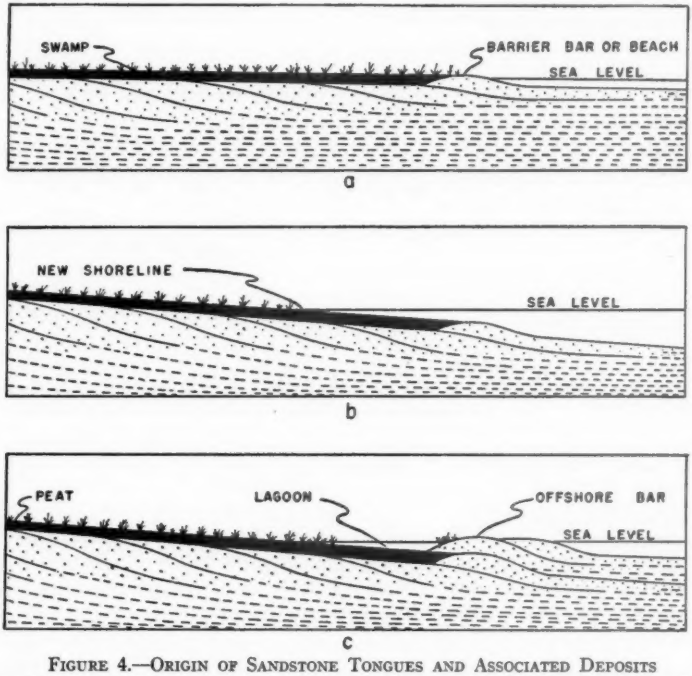

FIGURE 4.—ORIGIN OF SANDSTONE TONGUES AND ASSOCIATED DEPOSITS

a. Section showing probable conditions in shore zone. Swamp forms after growth of sandstone tongue has almost ceased.

b. Section showing same zone after sudden subsidence. Sea bottom has dropped about 50 feet, but last beach is only slightly submerged.

c. Section showing same zone with new sandstone tongue forming as offshore bar. Pre-existing barrier on old beach has determined where offshore bar will form.

cluding littoral marine sandstones. The transgressive differs from the regressive type in that its base is sharply defined while the top grades into the overlying shale. Some gradation was noted near the tips of some of the sandstone tongue in the Book Cliffs, but the majority retain their sharply defined upper surfaces until they finally face into the Mancos. The tips of the tongues should grade into the overlying as well as into the underlying shale, since they represent the subaqueous portion at the end of the regression. Thus upward gradation at the points would not be an indication of transgression.

All the littoral marine sandstone tongues of sidence which allowed rapid transgression of the sea followed by periods of quiet during which the shore line moved eastward. Spieker (1949, p. 65) postulates three causes for the regression of the sea: (1) uplift, (2) eustatic depression of the sea level, and (3) the feeding of sediment into a static basin. The first two are untenable since they both result in removal of previously deposited sediments. The body of a regressive sand would be eroded and reworked, leaving a deposit not much wider than the last beach.

The most plausible explanation is in the third situation where sea level remains constant as the beach builds seaward. The tops of sandstones formed in this way would be sharply

defined, while the base would grade downward in the shale. If the supply of sediment should diminish, swamps would form behind the last beach which would keep out the sea, or perhaps the vegetation alone would be sufficient barrier. It is also possible that bars formed in the shallow water offshore, later forming barriers behind which existed lagoons. This would, however, necessitate the filling of the lagoons before sizeable coal-forming swamps could exist, but in many cases no lagoonal deposits of creditable thickness exist between the sandstone and the overlying coal. A rapid subsidence of the marine basin would then allow the sea to sweep in and cover a large part of the lowlands along the coast. A new beach would form at the new shore line, and again regression would be instigated (Fig. 4).

Pike (1947, p. 15–19) states that rate of subsidence is the controlling factor in the intertonguing of the Upper Cretaceous of the Four Corners area. Subsidence of the basin undoubtedly occurred but probably in sharp pulses as previously mentioned. However, if it were assumed that slow subsidence accompanied the rapid pulses, the critical factor controlling the deposition would be the supply of sediment available. If the supply equals the space made available by subsidence, then there would be no shifting of environments and no intertonguing of facies. Such deposits may exist in the Book Cliffs but have not been identified. If the supply of sediment is less than the subsidence, then the result is a transgressive sandstone, of which there is no evidence in this area. Finally, if the supply exceeds the available space, regression of the shore line and the formation of regressive deposits will result. In this case the principal results would be (1) a higher angle between the time lines (bedding) and the body of the sandstone, and (2) more complete merger of the continental beds with the littoral marine sandstone, but there would always be a beach, and where it developed the sandstone would have a sharply defined top. The sandstone would rise stratigraphically seaward, the angle of rise depending on the rate of subsidence. If it is very slow the angle of rise may be small, however, with increased rate of sinking the supply-space ratio changes, and the angle may approach 90°. A still greater rate of subsidence would reverse the direction of rise and produce a transgressive deposit.

All the observed littoral marine sandstones of the Book Cliffs have sharply defined tops, grade downward into marine shale, and rise stratigraphically seaward at very low angles. Therefore, it follows that these sandstones formed as regressive beach deposits during periods of slow subsidence in a shallow basin in which supply was greatly dominant over space available.

Since each of the littoral marine sandstones of the Upper Cretaceous consists of a series of beaches built successively seaward, component parts of a beach are recognizable in each of the beds composing a sandstone tongue. Backshore, upper foreshore, lower foreshore, and near-shore laminae were noted. Eolian crossbedding can be found on the crests and lagoonal sides of some of the bars.

Fossils are not common in these littoral marine sandstones. Fucoids, worm tubes, and *Halymenites* are the most common evidence of life. In some localities *Ostrea* remains are numerous on the upper surfaces of the sandstones.

Offshore Bars

Similar to and of the same origin as the more extensive littoral marine sandstones are the offshore bar sandstones behind which the coal-bearing rocks formed. The pulses which formed them, however, were of lesser magnitude as shown by the fact that they have not, in many cases, affected the previously deposited coal-bearing rocks, but seem to have resulted in a new offshore bar being built on top of the previous one. In many cases the sea must have penetrated farther inland behind the bar, but few recognizable records are left. Apparently the increase of sediment available after these small subsidences allowed a rapid building of the offshore bars, usually rapid enough to exclude most of the marine water before any recognizable marine or littoral marine sediments could be deposited.

In many cases large coal swamps existed in these lagoonal and lowland areas, as indicated by thick coal deposits such as the Sunnyside and Kenilworth coals. The larger coal swamps did not flourish in these areas until the lagoons

were nearly filled with sediment, amounting in some cases to only a few inches or feet of fine clastics. Thus all the major coals of the Book Cliffs lie at horizons which correspond nearly exactly with the tops of the offshore bar sandstones. The exceptions may stem from a lack of information. Most theories of coal origin suggest bars protecting swamps from the sea, and excellent proof of this is supplied here.

The Castlegate "A", Aberdeen, Kenilworth, and Sunnyside coals of the Blackhawk, and some coals of the Price River, rest directly on the underlying sandstone or are separated from it by only a few inches or few feet of shale. These deposits must have formed in swamps which extended almost to the edge of the water. But they were probably separated from it by either a vegetative barrier or a small barrier beach. These barriers have not been definitely identified, but they must have existed to determine the position of the overlying bar sandstone which begins at the termination point of the basal coal unit. In the Spring Canyon member, for instance, the Hiawatha coal thins eastward and terminates in the cliff above the town of Helper, where the first offshore bar sandstone of the Spring Canyon member appears.

Perhaps the barrier bar was very small and was incorporated in the overlying offshore bar. The first well-defined bar sandstone extends a short distance into the lagoon and rests directly on the feather edge of the underlying coal. That these coals extended farther seaward and were eroded by the transgressing sea is discounted since the sandstone, beyond the coal termination, shows no trace of a white cap to indicate swamp water leaching. This absence is not due to erosion since the sandstones show no appreciable differences in thickness from the lagoonal side of the bar to the marine side, a distance of half a mile or more.

The regularity with which these offshore bars form above each other could be explained more easily if there were some indication of a hinge line at these points. It appears that, as soon as one lagoon was filled and a coal swamp had flourished, a subsidence flooded and killed the vegetation and caused a new bar sand to build on the previous one. This happened three times in the deposition of the Spring Canyon member, five times for the Kenilworth member, and at least five times for the Sunnyside member of the Blackhawk.

Modern barrier bars of this type are fairly common. Blanton (1951, p. 1425) believes that they are formed only on eastward-facing shore lines. If so, then the higher tides responsible for these bars may have also existed in the Late Cretaceous sea. This may account for the lack of such structures in the areas studied by Sears, Hunt, and Hendricks (1941), Pike (1947), Walton (1948), and others.

E. D. McKee (1951, personal communication) finds in a study of the barrier bars and lagoons of the Corpus Christi area a modern situation which closely resembles that of the Upper Cretaceous rocks of the Book Cliffs. He has found types of cross lamination recognized in Upper Cretaceous sandstones, and sand dunes which are driven lagoon-ward by the winds of the Gulf. The lagoonal deposits of both locations are similar, although those of Corpus Christi lack the thick plant accumulations of the Upper Cretaceous.

Coal-Bearing Rocks

The coal-bearing rocks comprise the rocks of the lagoons, estuaries, flood plains, swamps, and lowlands, referred to as lagoonal deposits. They consist of buff to gray sandstone lenses and gray shale with some coal beds. The sediments range from medium-grained sandstone to shales and include some fireclay.

Some portions of the coal-bearing rocks are evenly bedded and show varvelike bedding; some show almost no bedding, while others exhibit deltaic structure. The high organic content of all these rocks causes them to crumble rapidly on exposure to weathering. It might also be the cause of rapid oxidation found where these sediments have been associated with burning coals resulting in bright reds and yellows.

Marine shale tongues not yet recognized must be present in a few places in these deposits behind the barrier bars. A few thin marine shale tongues in the Sego member bear sharks' teeth (*Lamna* sp.).

The coal deposits, which correspond to the tops of barrier bars, probably were formed in fresh-water swamps. No fossils have been found, but many brackish-water species of *Corbula*, *Corbicula*, *Modiola*, *Anomia*, and *Ostrea* were found a few inches above the Kenilworth and Upper Sunnyside coal beds. This may indicate deposition in brackish waters, but probably these brackish-water forms moved in after the marine flood which killed off the coal-forming plants. In some cases two or more coals lie directly on one another or are separated by only a few inches of bone. This would indicate that the marine transgressions lasted only long enough to kill off the vegetation, and with recession or exclusion of marine waters, a new swamp equivalent in area to the previous one formed. It may also mean that the coal swamps were inundated near the sea but continued to flourish farther inland where the coals are thicker.

The lateral extent of this coal-bearing environment is not known but must have been hundreds of miles. The transverse extent of individual swamps varied greatly but probably averaged 50 miles or more. Probably none of the coal zones is continuous for such distances, but the Anchor coal (redefined) is reported by Fisher (1936, p. 46) and Erdmann (1934, p. 81) to be traceable for nearly 100 miles from west to east. However, such widespread tracing is unreliable since one burned outcrop, of which there are many in the Book Cliffs, could easily obscure stratigraphic relations. This is particularly true in the lagoonal areas where the coal does not rest on indicative massive white-capped sandstones.

The purity of the coals of this region has been discussed by Clark (1928, p. 81–93), Spieker (1931, p. 68), Erdmann (1934, p. 88–92), and D. J. Fisher (1936, p. 47–55), but remains difficult to explain. Some sediment must have been still entering the near-by sea, and winds must have been active to produce the eolian deposits.

The supply of sand available for the formation of the seaward portions of the sandstone tongues must have been considerably less than for the much thicker landward portions (Pl. 3). This suggests that either erosion or transportation had become inadequate to maintain the thickness. Whatever the cause, the seaward growth of each sandstone tongue must have almost ceased before the next pulse of subsidence, as shown by their gradual thinning and termination (Pl. 3).

Calcareous concretions at the ends of the tongues may indicate fairly long periods of little deposition (Spieker, 1949, p. 68). If so, then the main coal-forming swamps must have existed during these periods. Since the coals are relatively free of inorganic material and since the calcareous concretions indicate little deposition, this must have been a time of almost no erosion or transportation of debris.

Inland Flood-Plain Rocks

The lithology of the rocks of the inland facies has been discussed in the description of the Price River and North Horn formations. They closely resemble the fresh- and brackish-water deposits of the lagoonal facies except that coal beds are thin and rare.

Fresh-water fossils have been reported by Spieker. *Unio*, *Sphaerium*, *Viviparus*, *Goniobasis*, and *Physa* are common, and vertebrate remains are of local importance.

Cyclic Nature of Units

The rocks of the Book Cliffs present a clear picture of cyclic deposition. Plate 3 shows the many repetitions in the stratigraphic sequence, from the uppermost part of the Mancos shale upward through littoral marine sandstone, into coal-bearing rocks, and again into marine shale; or from any point westward along a given time line through littoral marine sandstone into coal-bearing deposits.

Cycles of different magnitudes are represented in Plate 3. There are many small cycles which can be grouped into larger cycles (megacycles) according to the extent of the littoral marine sandstones. All the cyclothems (deposits of a cycle) should, in theory, have the following sequence of units:

cyclothem
4. coal
3. lagoonal deposits
2. littoral marine sandstone
1. marine shale

Most of the littoral marine sandstone units are offshore bar sandstones and do not usually extend far into the lagoonal deposits. However, at certain horizons, they are much more extensive, and these constitute the basal units of the various members of the Blackhawk and Price River formations. Each consists of deposits of many small cycles, which together comprise the deposits of one megacycle. The cyclic nature of these rocks suggested the subdivision of the Upper Cretaceous formations. Since there are six megacycles in the Blackhawk, six members have been distinguished. In the portion of the Price River which lends itself to subdivision, four members have been named.

The number of cycles in a megacycle must vary. The exact number in any megacycle is difficult to determine, because the coal-bearing rocks are not so well exposed as the littoral marine sandstones. Therefore no attempt was made to subdivide further the lagoonal deposits. Except for the basal cyclothem of each megacyclic deposit, the littoral marine sandstone tongue and probably the marine shale of most cyclothems are absent in the lagoonal area, leaving two units to be distinguished—the lagoonal and coal units.

At the end of each cycle, which is a series of changes instigated by a sharp pulse of basin subsidence, there probably would be a period of very little clastic sedimentation, as attested to by the presence of thin limestone stringers and zones of limestone nodules in the marine shale opposite the tips of the major sandstone tongues. During these periods of quiet, coal swamps flourished in the coastal area and persisted until the next pulse of subsidence. The last coal bed of a cycle should, therefore, be thicker and more continuous than the beds formed during the filling of the lagoon.

In the Spring Canyon member of the Blackhawk, three thick coals are present as well as several others of lesser thickness and extent. These three coals lie at horizons which correspond to the tops of offshore bar sandstones in the vicinity of Helper (Pl. 3). At least five cycles are represented in the Spring Canyon megacycle. The probable sequence in this megacycle near the seaward edge of the lagoonal deposits (just west of Helper) is:

Spring Canyon cycle (megacyclothem)	4. coal 3. lagoonal beds 4. coal 3. lagoonal beds 4. coal (Hiawatha) 2. sandstone 1. shale 2. sandstone 1. shale 2. sandstone 1. shale

Farther inland the lagoonal unit drops out, leaving the thin and discontinuous coal unit. It may, in some instances, extend a considerable distance inland and be associated with deposits of the inland facies.

One gratifying result of this stratigraphic study has been the visualization of cyclothems which have obviously originated in the natural fashion of oscillation, as all cycles should operate. This is opposed to the implication, in consideration of Pennsylvanian cycles, of widespread and simultaneous changes. The type of cycle which produced the cyclothems here described was one which involved the shifting of sedimentary environments in response to a pulse of basin subsidence followed by gradually declining clastic deposition. The evidence found in the Book Cliffs thus bears directly on the problem of cyclic deposition.

References Cited

Blanton, S. L., Jr., 1951, Origin of offshore bars (Abstract): Geol. Soc. America Bull., v. 62, p. 1425.

Caster, Kenneth E., 1934, The stratigraphy and paleontology of northwestern Pennsylvania: Bull. Am. Paleontology, v. 21, p. 1–185.

Clark, Frank R., 1928, Economic geology of the Castlegate, Wellington, and Sunnyside quadrangles, Carbon County, Utah: U. S. Geol. Survey Bull. 793, 162 p.

Cobban, William A., and Reeside, J. B., Jr., 1952, Correlation of the Cretaceous formations of the western interior of the United States: Geol. Soc. America, Bull., v. 63, p. 1011–1044.

Cross, W., and Purington, C. W., 1899, Description of the Telluride quadrangle, (Colorado): U. S. Geol. Survey Geol. Atlas, folio 57, p. 4.

Erdmann, C. E., 1934, The Book Cliffs coal field in Garfield and Mesa counties, Colorado: U. S. Geol. Survey, Bull. 851, 145 p.

Fisher, A. G., 1954, Arbitrary cut-off in stra-

tigraphy: Am. Assoc. Petroleum Geologists Bull., v. 38, p. 926–928.

Fisher, D. J., 1936, The Book Cliffs coal field in Emery and Grand counties, Utah: U. S. Geol. Survey Bull. 852, 102 p.

Lee, W. T., 1912, Coal fields of Grand Mesa and the West Elk Mountains, Colorado: U. S. Geol. Survey Bull. 510, 237 p.

Pike, W. S., 1947, Intertonguing marine and nonmarine Upper Cretaceous deposits of New Mexico, Arizona, and southwestern Colorado: Geol. Soc. America Mem. 24, 100 p.

Sears, J. D., Hunt, C. B., and Hendricks, T. A., 1941, Transgressive and regressive Cretaceous deposits in southern San Juan Basin, New Mexico: U. S. Geol. Survey Prof. Paper 193, 19 p.

Spieker, E. M., 1931, The Wasatch Plateau coal fields, Utah: U. S. Geol. Survey Bull. 819, 206 p.

— 1946, Late Mesozoic and Early Cenozoic history of central Utah: U. S. Geol. Survey Prof. Paper 205-D, 43 p.

—— 1949, Sedimentary facies and associated diastrophism in the Upper Cretaceous of central and eastern Utah: Geol. Soc. America Mem. 39, p. 55–81.

Spieker, E. M., and Reeside, J. B., Jr., 1925, Cretaceous and Tertiary formations of the Wasatch Plateau, Utah: Geol. Soc. America Bull., v. 36, p. 429–454.

Walton, P. T., 1944, Geology of the Cretaceous of the Uinta Basin, Utah: Geol. Soc. America Bull., v. 55, p. 91–130.

Wheeler, Harry E., and Mallory, V. Standish, 1953, Designation of stratigraphic units: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 2407–2421.

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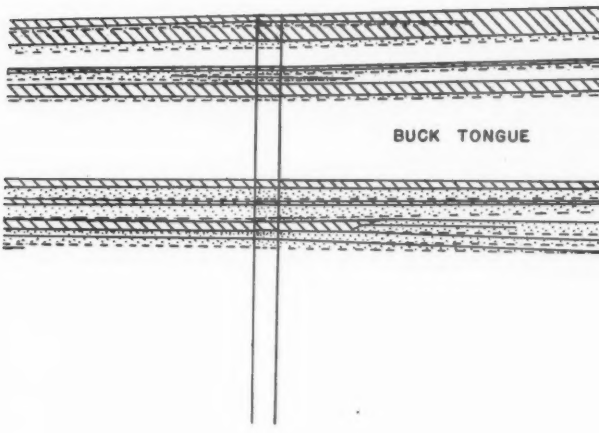
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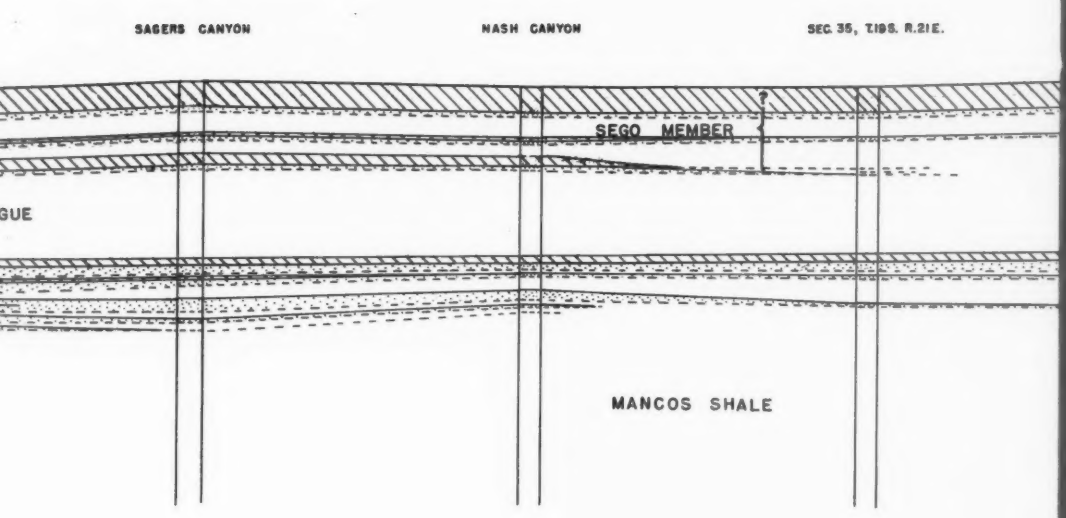
SAGERS CANYON
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SEC. 35, T.19S. R.21E.
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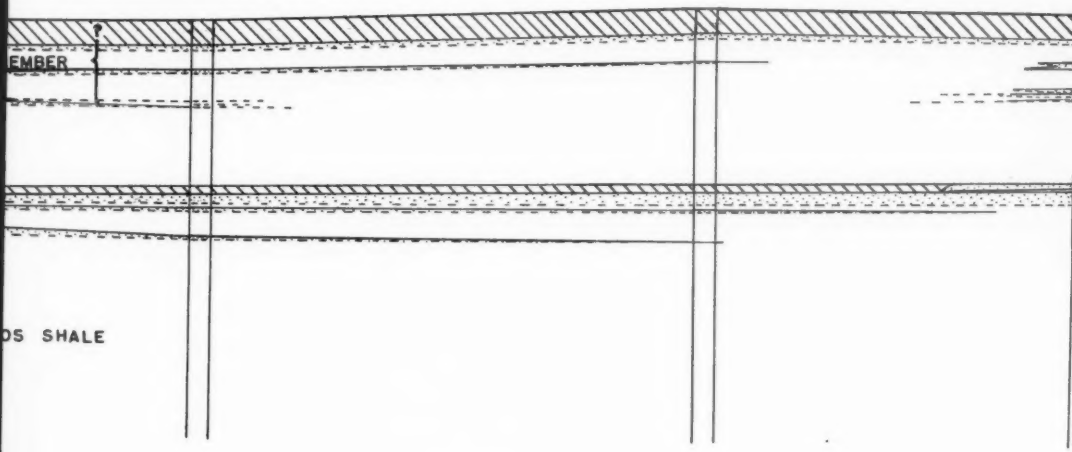
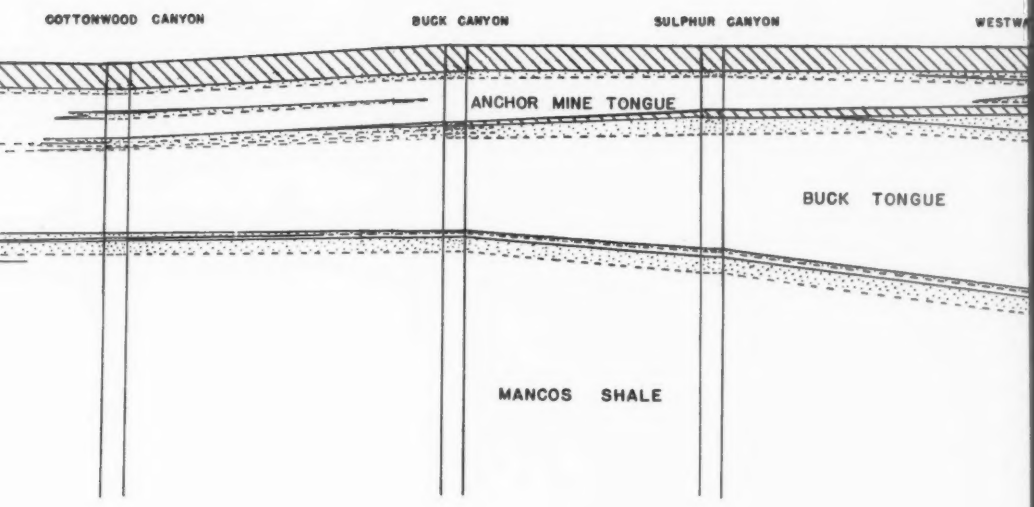
MANCOS SHALE

SEC. 35, T.10S. R.21E.

SEC.1, T.20S., R.22E.

COTTON





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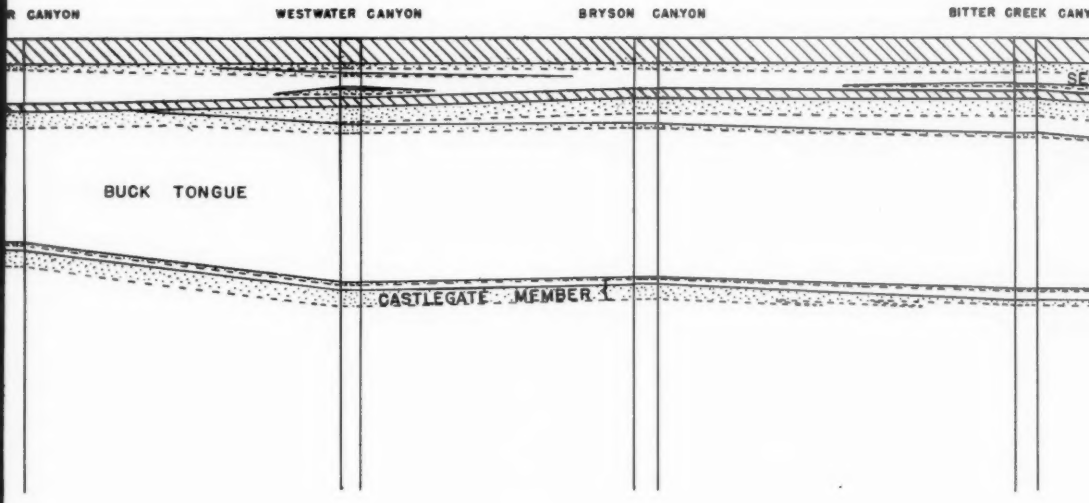

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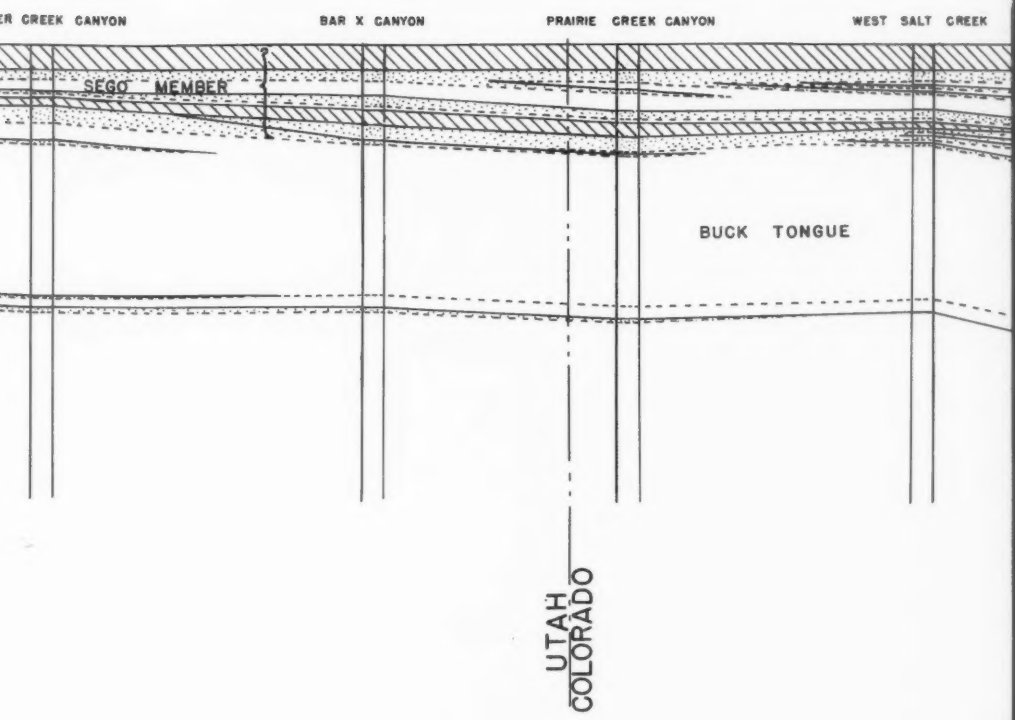
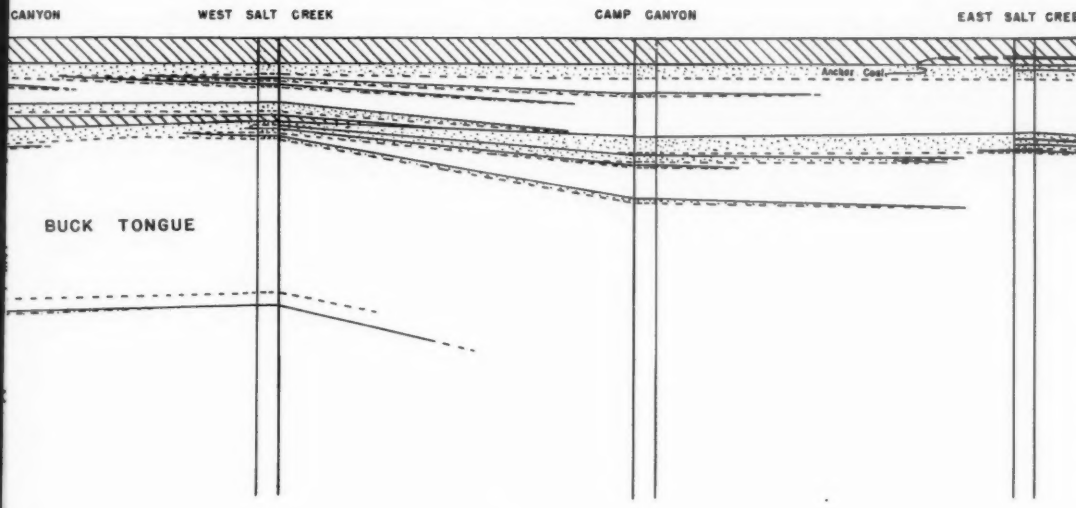


BOOK CLIFFS BETWEEN STORRS, UTAH, AND GRAND J



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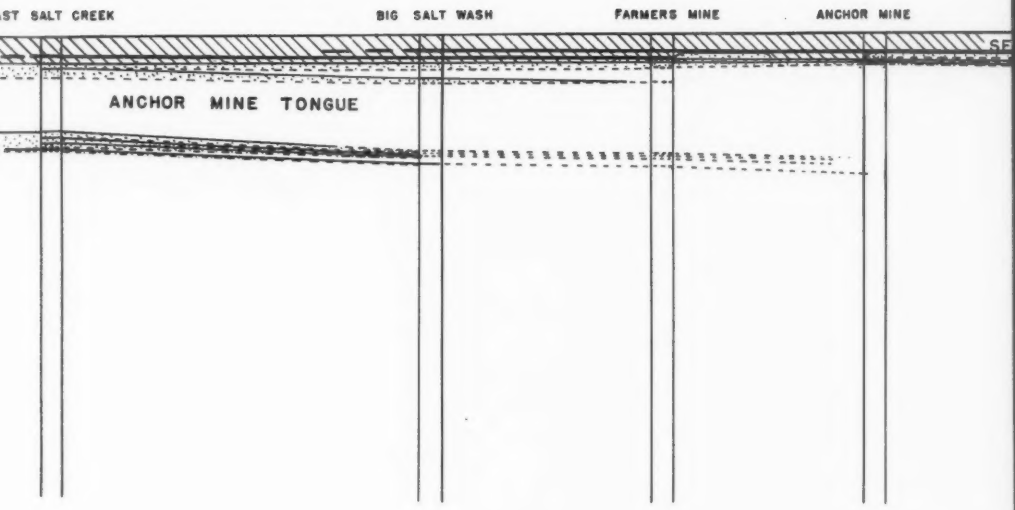
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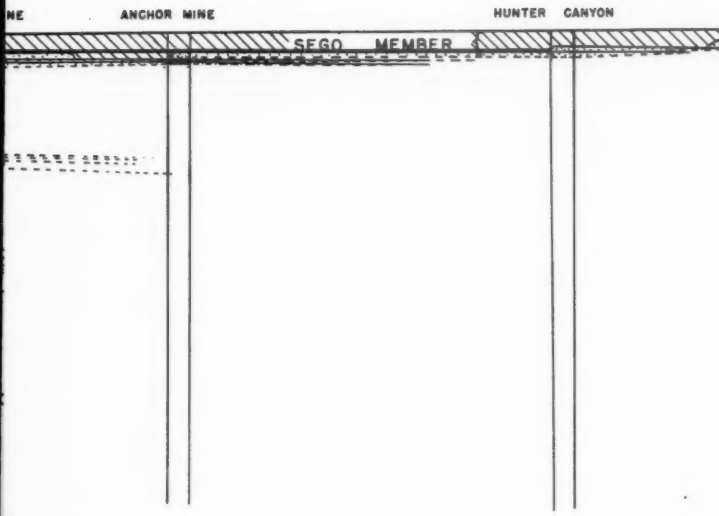
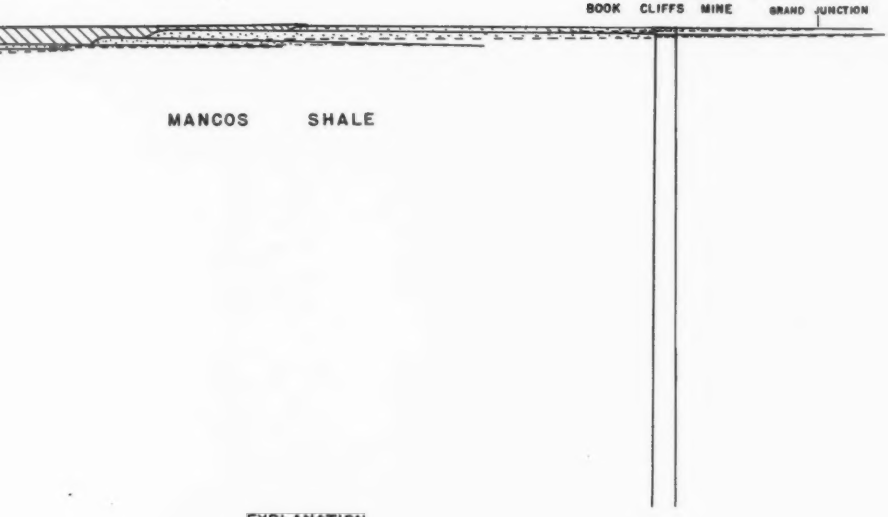


FEATURES O

PRINCIPAL C

WHITE CAP

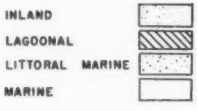
Young, Pl. 3 *(Right Half)*

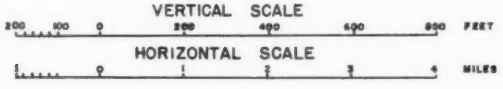


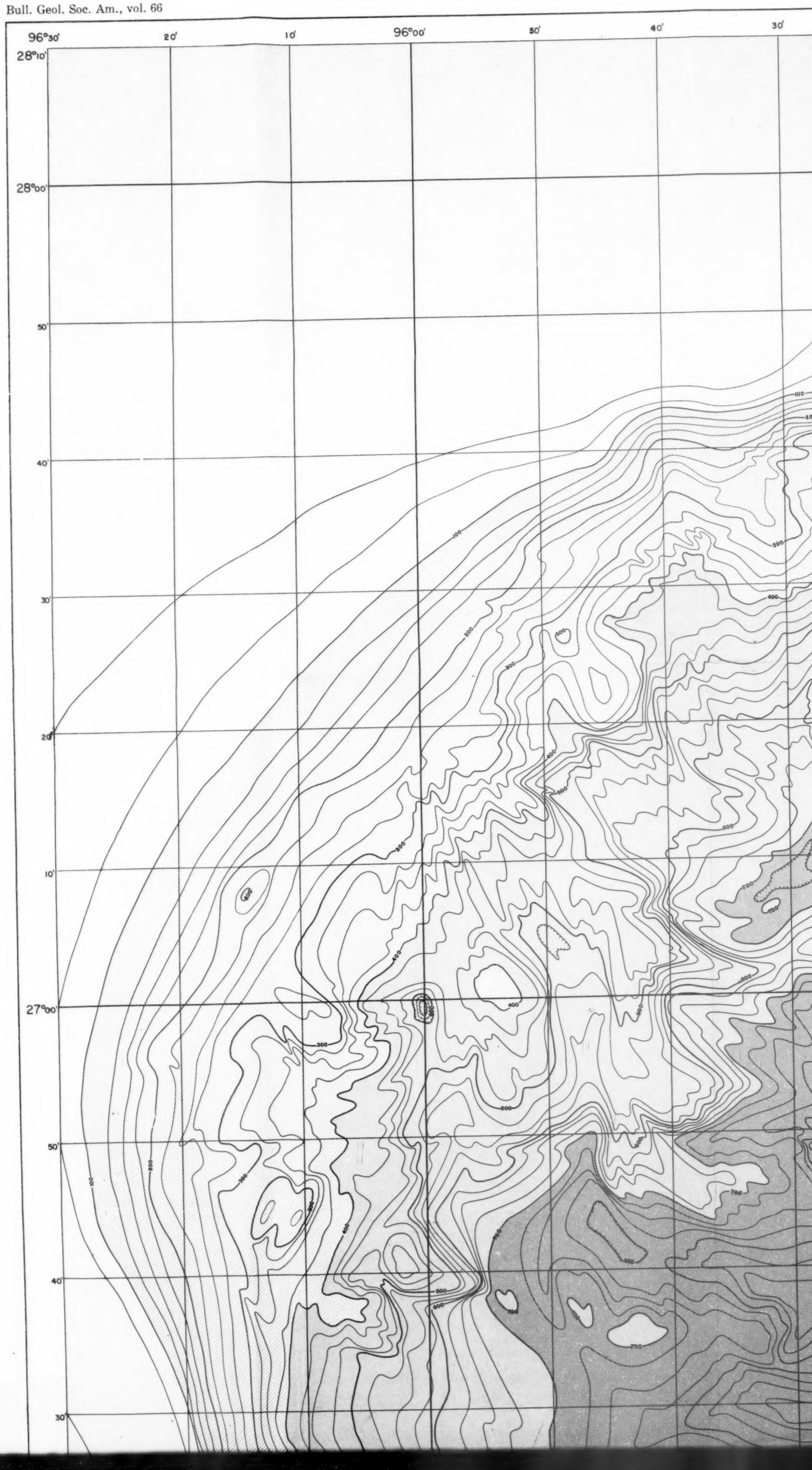
EXPLANATION

FEATURES OF SANDSTONES

FACIES



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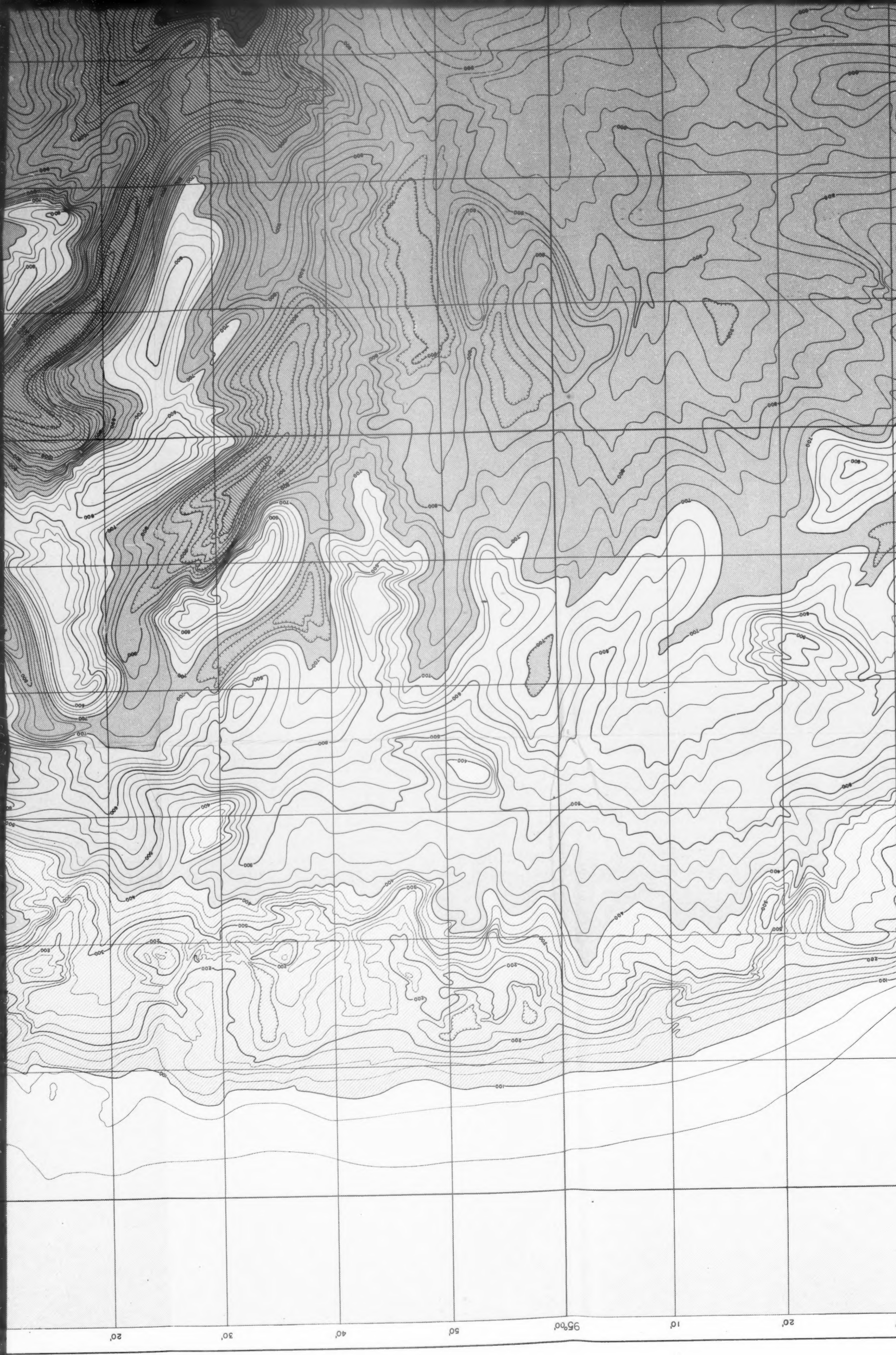
20'
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20' 10' 94°00' 50' 40' 30' 20'

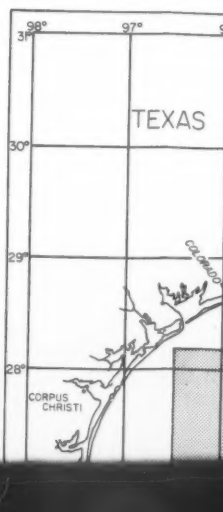
DEPTH IN FATHOMS

0 to 100

100 to 400

400 to 700

700 to 1000



10' 93°00' 50' 40' 30' 92°20'

28°10'

28°00'

50'

40'

30'

20'

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27°00'

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96° 95° 94° 93° 92° 91° 90° 89° 88°

31°

30°

29°

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LOUISIANA

MISS.

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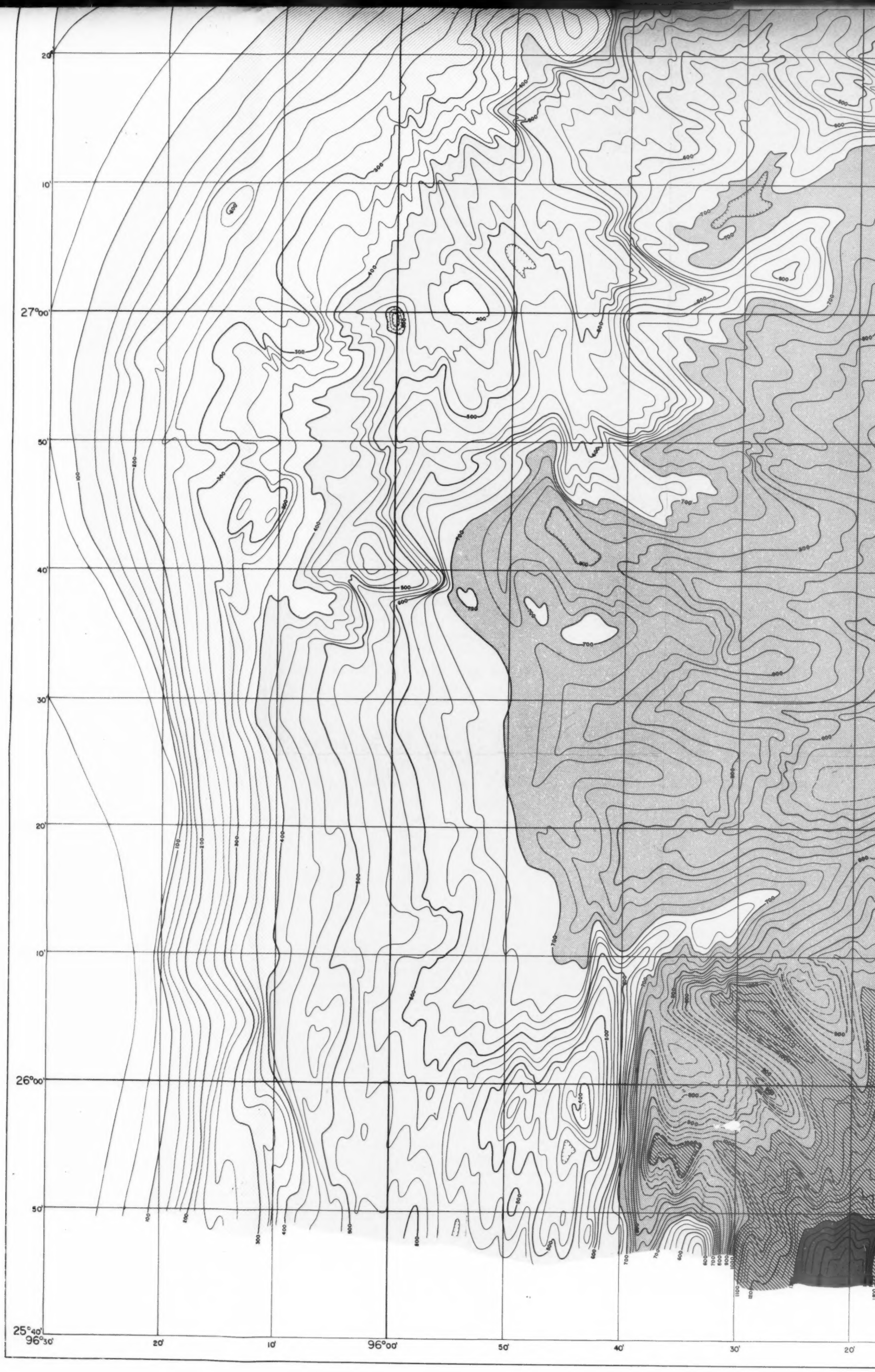
EASTERN PART

PLATE 2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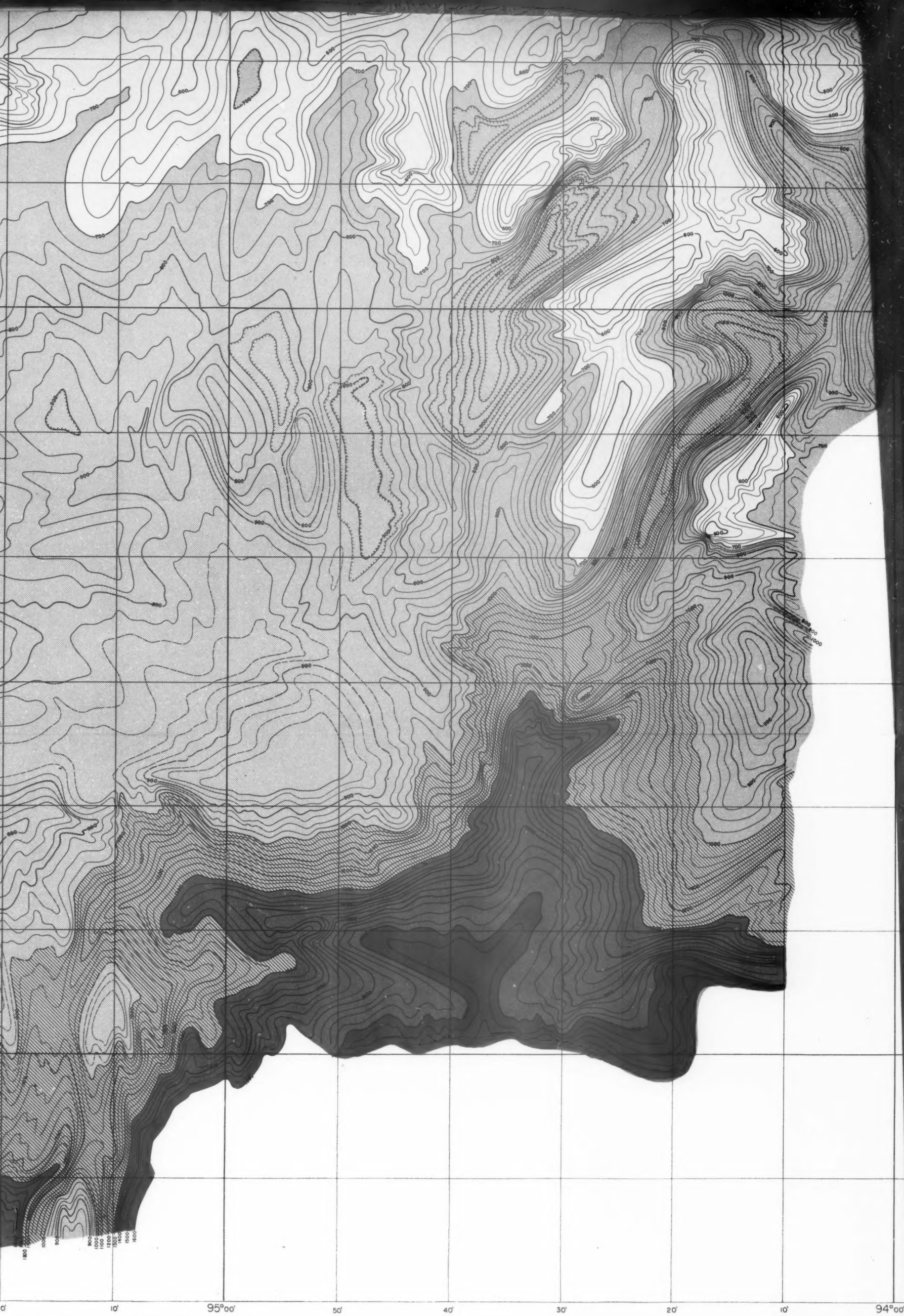
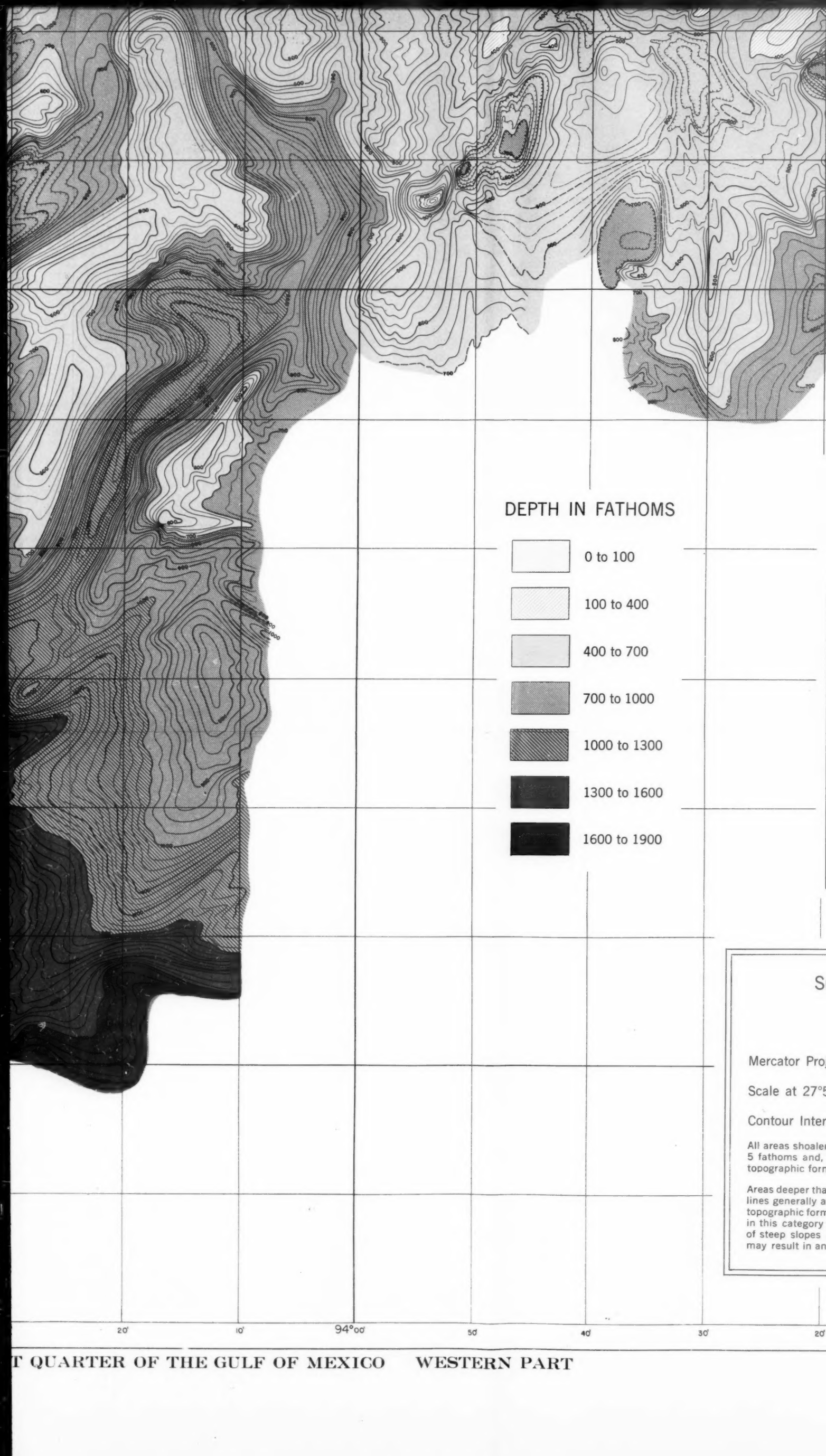
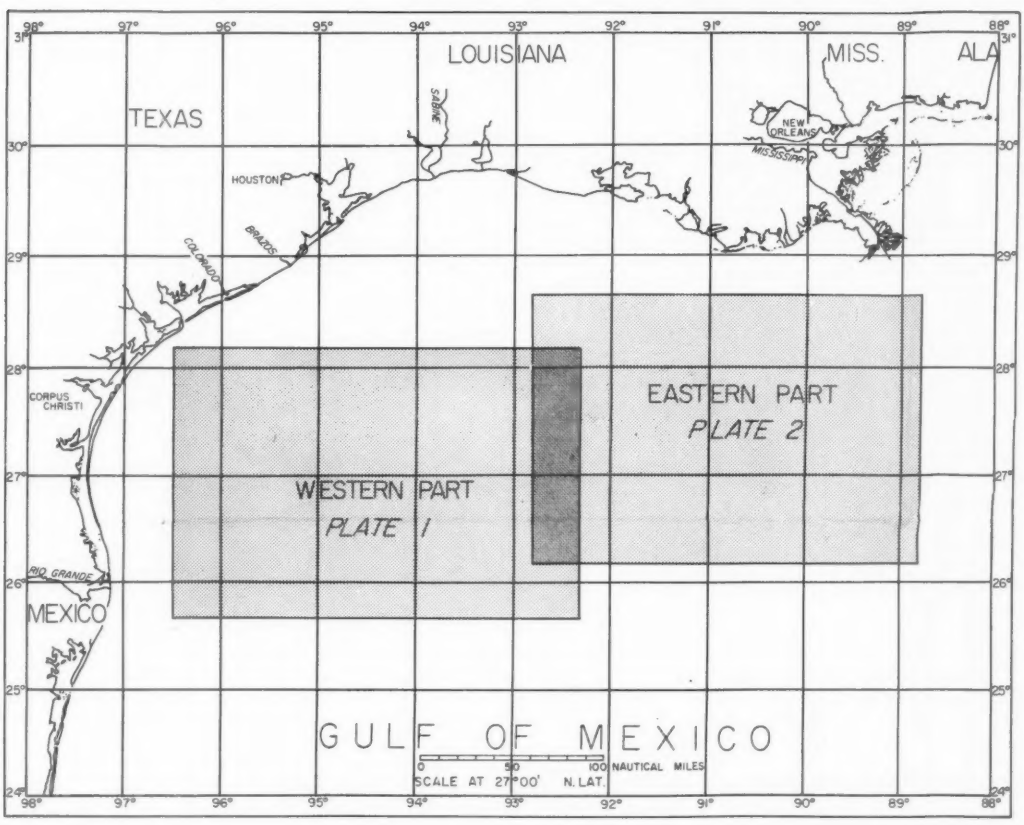


SUBMARINE TOPOGRAPHY IN THE NORTHWEST QUARTER OF THE GULF OF MEXIC



T QUARTER OF THE GULF OF MEXICO WESTERN PART



Submarine Topography in the Northwest Quarter of the GULF OF MEXICO

WESTERN PART

Projection

27°50′ N. Lat. = 1:500,000

Interval = 25 fathoms

Data from U. S. Coast and Geodetic Survey Hydrographic Surveys

Contouring Sponsored by Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

Contoured by Betty Lee Gealy, December, 1950

noaler than 250 fathoms are accurate to within and, in most cases, to within 1 fathom. The c form is essentially as indicated.

er than 250 fathoms contoured in solid contour ally are accurate to within 25 fathoms, and the c form is essentially as indicated. Also included egory are two other groups: (1) In some areas opes and high relief, small errors in position in an error in depth as great as 100 fathoms, although the topographic form is essentially as indicated. (2) In some areas of very low relief and gentle slopes, the topographic form may not be exactly as indicated although the error in depth will be less than 25 fathoms.

Areas contoured in dashed contour lines include topographic forms which are in doubt and the error in depth may exceed 100 fathoms in some cases. The topographic forms indicated are considered to be the most probable of several interpretations.

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TOPOGRAPHY OF THE CONTINENTAL SLOPE IN NORTHWEST GULF OF MEXICO

By Betty Lee Gealy

Abstract

Topographic charts of the continental slope in the northwest Gulf of Mexico were constructed largely from soundings from unpublished hydrographic surveys of the U. S. Coast and Geodetic Survey. The configuration of the slope and the nature of its sediments suggest a genetic interpretation for its varied topography and high relief.

Twenty-one profiles show that the break in slope at the edge of the continental shelf is marked only by a gradual steepening. A hummocky zone on the upper part of the slope shows features resembling landslide scars and deposits, sediment-flow scars, and surface expressions of faults. These features indicate that the upper part of the slope is underlain by a mass of relatively unstable sediments, probably deposited during the Pleistocene when sea level was near the present 75-fathom curve.

Steep slopes and a lack of hummocks on the lower part of the slope indicate that most of the underlying sediments are relatively stable. Many steep-sided troughs cut this portion of the slope. Their position and configuration suggest a history which may involve (1) subaerial erosion, (2) submergence and tilting, (3) masking of the upper part of the slope by Pleistocene sediments, and (4) extensive modification by the failure of sediment under the stress of gravity.

The Sigsbee Scarp, at the base of the slope, is at least 340 miles long and as high as 4800 feet. It may represent one of a series of large faults surrounding the Gulf of Mexico and downthrown toward the Sigsbee Deep.

The Sigsbee Deep is a basin of gentle slopes and subdued relief extending from the base of the Sigsbee Scarp, at 1700 fathoms, to a maximum depth of about 2200 fathoms.

Although salt domes may underlie a number of features of low relief on and near the edge of the continental shelf, there is no evidence which indicates they have been an important geomorphic factor on the continental slope.

The topography and sediments suggest that in the northwest Gulf of Mexico there is a large mass of sediment under the stress of gravity and failing by landslides, faults, creep, sediment flows, and intermediate types of failure.

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Introduction

The topography of the continental slope in the northwest Gulf of Mexico has remained virtually unknown until recent years, and neither adequate topographic charts of the region nor accurate descriptions of the topography are available. In 1949 the writer began: (1) to compile and contour the most accurate soundings available in the region, (2) to study the factors which may have influenced the development of the topography on the continental slope, and (3) to seek a genetic interpretation of the topography.

The U. S. Coast and Geodetic Survey completed a series of hydrographic surveys in 1940 using echo-sounding devices and good positional control. A few of the data obtained on these surveys were published on U. S. Coast and Geodetic Survey Charts numbers 1116 (1943) and 1117 (1941) (scale approximately 1:500,000). Random soundings taken from the hydrographic surveys are shown on these charts numerically, and 100-fathom curves are shown on the continental slope to a maximum depth of 500 fathoms. However, most of the data obtained on these surveys remains unpublished on file in the offices of the Survey in Washington, D. C. While mechanisms have been devised recently which permit even more precise depth and positional control than that obtained on these surveys, this region probably will not be resurveyed in detail with these new devices for many years. Echo-sounding records made since 1945 aboard the Research Vessel Atlantis of Woods Hole Oceanographic Institution and aboard surveying vessels of the Hydrographic Office of the U. S. Navy have provided additional unpublished soundings.

The continental slope has been found to have a varied topography of high relief which does not resemble the topography on the continental slopes of either the Atlantic or Pacific coast of North America. The topography, the agencies operating in the Gulf, and the sediments underlying the region suggest several working hypotheses concerning the origin of the topography.

Acknowledgments

The writer is grateful to Mr. H. C. Stetson for encouragement and advice throughout the preparation of the accompanying topographic charts and manuscript.

Woods Hole Oceanographic Institution has given financial assistance to the writer during the summers of 1949–1952, and has provided necessary equipment and paid expenses throughout the project.

The National Science Foundation has furnished financial assistance through a Predoctoral Fellowship for the academic year 1952–1953.

The writer is indebted to members of the U. S. Coast and Geodetic Survey for their enthusiastic co-operation in the preparation of the topographic charts. Copies of the original unpublished hydrographic surveys, on file in the offices of the Survey, as well as detailed information concerning the positional control of those surveys were made available to the writer. Members of the Survey spent much time discussing the problems and limitations of marine surveying and methods of compiling and correcting data. In addition, members read the section of the manuscript entitled "Construction of Topographic Charts" and offered a

number of helpful suggestions. Adm. L. O. Colbert, Adm. Robert W. Knox, Mr. H. R. Edmonston, Comdr. F. B. Quinn, and Mr. R. H. Carstens were especially helpful.

The Hydrographic Office of the U. S. Navy provided echo-sounding records taken in the northwest Gulf of Mexico prior to 1948. The U. S. Geological Survey provided a Salzman Projector which was used to reduce and compile base charts. Dr. W. H. Bradley and Mr. K. P. Cartrette were of particular assistance.

Dr. S. D. Wilson has read the portions of the manuscript dealing with the behavior of sediments under stress and has suggested revisions. Dr. M. P. Billings, Dr. Francis Birch, Dr. L. D. Leet, Dr. Charles E. Stearns, and Dr. P. D. Trask have read the manuscript and offered many helpful suggestions. Discussions with the late Mr. M. G. Cheney, Dr. H. N. Fisk, Dr. F. P. Shepard, and Adm. Paul A. Smith have helped the writer in the formation of ideas. Miss Jean Stilwell constructed the profiles used in Figures 2, 3, and 4.

Construction of Topographic Charts

Location and Extent

Topographic charts of the continental slope in the northwest Gulf of Mexico have been constructed (Pls. 1, 2). The contoured area is bounded on the north and west by the edge of the continental shelf at about the 75-fathom curve, on the south by the southern limit of the hydrographic surveys of the U. S. Coast and Geodetic Survey in the northwest Gulf of Mexico, and on the east by the Mississippi Trough southwest of the Southwest Pass of the Mississippi River. The area extends nearly 400 miles[1] from east to west and more than 75 miles from north to south. The shoreward margin of the contoured area is about 45 miles from the mouth of the Rio Grande, 120 miles from the mouth of the Sabine River, and 50 miles from shore west of Southwest Pass.

Sources of Data

Most soundings used in contouring the area were taken from original unpublished hydrographic surveys of the U. S. Coast and Geodetic Survey. Photostats or bromide copies of these surveys can be obtained at cost from the office of the Survey in Washington, D. C. by those interested in seeing the original material. The hydrographic surveys used in constructing the charts are given in Table 1. Copies of the hydrographic surveys were photographed on a common scale of 1:120,000 and were used as base maps for contouring.

Table 1.—Hydrographic Surveys of the U. S. Coast and Geodetic Survey Used in Preparing Plates 1 and 2

Survey number	Original scale	Date of survey
H-5768	1:120,000	1934
H-6184	1:80,000	1936
H-6185	1:80,000	1936
H-6292	1:80,000	1937
H-6293	1:80,000	1937
H-6499-A	1:240,000	1939
H-6499-B	1:80,000	1939
H-6501	1:120,000	1939
H-6502	1:120,000	1940
H-6546	1:120,000	1940
H-6547	1:160,000	1940

In addition to the above soundings, echo-sounding records made aboard the Research Vessel Atlantis of Woods Hole Oceanographic Institution, and aboard surveying vessels of the Hydrographic Office of the U. S. Navy were available. However, as positional control is relatively poor for these records, the soundings from them were used only to validate the existence of large topographic features in areas not covered adequately by U. S. Coast and Geodetic Survey soundings. These records also were used to outline what the writer has termed the Sigsbee Scarp (Fig. 1).

Construction of the charts was completed in September 1950, and data acquired since then have not been incorporated.

Contouring Technique

The techniques and limitations of contouring submarine soundings have been discussed thoroughly by Veatch and Smith (1939), who state (p. 84), "In any approach, the fundamen-

[1] Throughout, the term miles refers to nautical miles.

tal problem is to fit a type of topography to the facts—the soundings," and (p. 83), "... the stream erosion type as delineated is, in the light of our present knowledge of the continental shelf, the most satisfactory topography possible." Thus, where soundings are scanty, Veatch and Smith have superimposed a detailed subaerial erosion pattern even where it is not indicated clearly.

On Plates 1 and 2 detailed topography is omitted except where soundings are dense. As a result, topographic features are drawn smoother and more regular than they probably are. Preconceived topographic types have been avoided as much as possible as the origin of some submarine topographic features is still under study, and charts like these are used in such studies. One disadvantage of this technique is that the topography appears to differ in complexity from place to place whereas this is due only to variation in concentrations of soundings.

Areas shoaler than 250 fathoms were contoured originally with a 5-fathom contour interval, and those deeper than 250 fathoms with a 25-fathom interval.

The separate base maps were compiled, re-

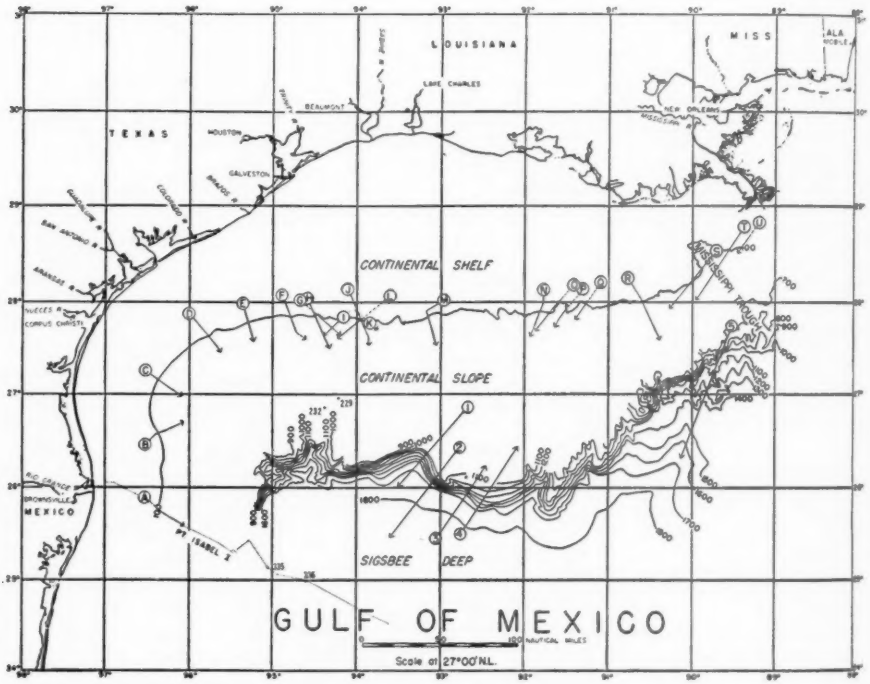


FIGURE 1.—MAP OF SIGSBEE SCARP

Depths in Fathoms. Contour interval 100 fathoms.

Location and extent of Sigsbee Scarp indicated by contour lines from 700 fathoms to 1800 fathoms.

Lines *A* through *U* are locations of profiles across break in slope shown on Figures 2, 3, and 4. Arrows point to right end of profiles. Sources of echo-sounding records used in constructing profiles given in Appendix 1. Dotted portions taken from Plates 1 and 2.

Lines *1* through *5* are locations of echo-sounding records crossing Sigsbee Scarp shown on Figure 5. Arrows point to right end of echo-sounding records.

Dotted line shows location of Port Isabel I. sedimentary profile shown on Figure 6. Traverse made by ATLANTIS in 1947.

Long cores 335 and 336 were taken by ATLANTIS in 1947 and include a sand. Cores 229 and 232 taken in submarine trough by ATLANTIS in 1951.

duced, and transferred onto a Mercator Projection with a scale of 1:250,000 at 27°30′ N. Lat. A contour interval of 25 fathoms was used throughout. This single chart was then reduced 50 per cent and reproduced as Plates 1 and 2.

Sources of Error

Positional control.—Positional control on the hydrographic surveys of the U. S. Coast and Geodetic Survey on the continental shelf is excellent. Sono-radio buoys were used, and the precise distances between buoys was established in many cases by taut-wire surveys. On charts having a scale of 1:500,000 (Pls. 1, 2), no error should be evident.

Positional control on the continental slope is less accurate. Sono-radio buoys were established on shoals at about the 75-fathom curve, and, under the most favorable conditions, positions accurate to one part in 200 can be obtained up to 100 miles from the buoys. Beyond the range of the buoys, lines were fixed only by dead reckoning and celestial navigation.

Crossings of sounding lines are checks on the positional control, and some sounding lines were adjusted slightly to make the soundings at the crossings match. Even on the deepest part of the continental slope mapped, the maximum error in the position of a sounding probably is no greater than 1 mile. Throughout most of the map, the error is less than half a mile which is small for marine navigation.

Depth control.—The U. S. Coast and Geodetic Survey soundings were taken with visually read echo soundings which in some cases are less reliable than those automatically recorded, but which are infinitely more accurate than the older wire soundings. These soundings were corrected for temperature, salinity, and pressure, but no corrections for the slope of the ocean bottom were made. It can be assumed that all slopes are slightly steeper and that all valleys are somewhat deeper than indicated on the charts, although, at the published scale, these errors should be small.

Frequency of observations.—As time must be allowed for the return of the echo, fewer soundings per minute can be taken in deep water than in shallow. Thus, assuming the speed of the ship is 12 knots, and the speed of sound in water is 800 fathoms per second, a maximum of 1200 soundings per mile can be recorded at 100 fathoms, 120 soundings per mile at 1000 fathoms, and 60 soundings per mile at 2000 fathoms. However, as these soundings had to be read visually, only about 6 soundings per mile were recorded at 100 fathoms, and 2 soundings per mile between 1000 and 2000 fathoms. From the point of view of the cartographer it is regrettable that most sounding lines are perpendicular to the edge of the continental shelf and that few are parallel to the edge of the shelf. This plan has influenced the accuracy of the contour lines considerably and, in some cases, has been the sole reason for placing an area in the third category of accuracy (Pls. 1, 2).

On the continental shelf, the sounding lines are 2–5 miles apart and on deeper parts of the continental slope the lines are 5–10 miles apart.

Topography

Continental Shelf

The continental shelf (Fig. 1) in the northwest Gulf of Mexico extends from shore to the break in slope at a depth of about 75 fathoms. Off the mouth of the Rio Grande the shelf is about 45 miles wide and has an average gradient of about 2 fathoms per mile (1:500); at 93° 30′ W. Long. it is 120 miles wide and has an average gradient of less than 1 fathom per mile (1:1000). West of the Mississippi Trough (Fig. 1) the shelf is about 60 miles wide and has an average gradient of about 1.5 fathoms per mile (1:700). Topographic irregularities on the shelf have a relief of only a few feet or tens of feet, whereas those on the continental slope have a relief of hundreds or thousands of feet.

The break in slope at the edge of the continental shelf must be chosen arbitrarily as it is marked only by a gradual increase in gradient in most areas. In some cases topographic irregularities, such as bioherms, further obscure the break. Profiles have been made from the available echo-sounding records that cross the 100-fathom curve at nearly right angles. The profiles are shown on Figures 2, 3, and 4, and the locations of the profiles on Figure 1. The "abrupt shelf-break" in the northwest

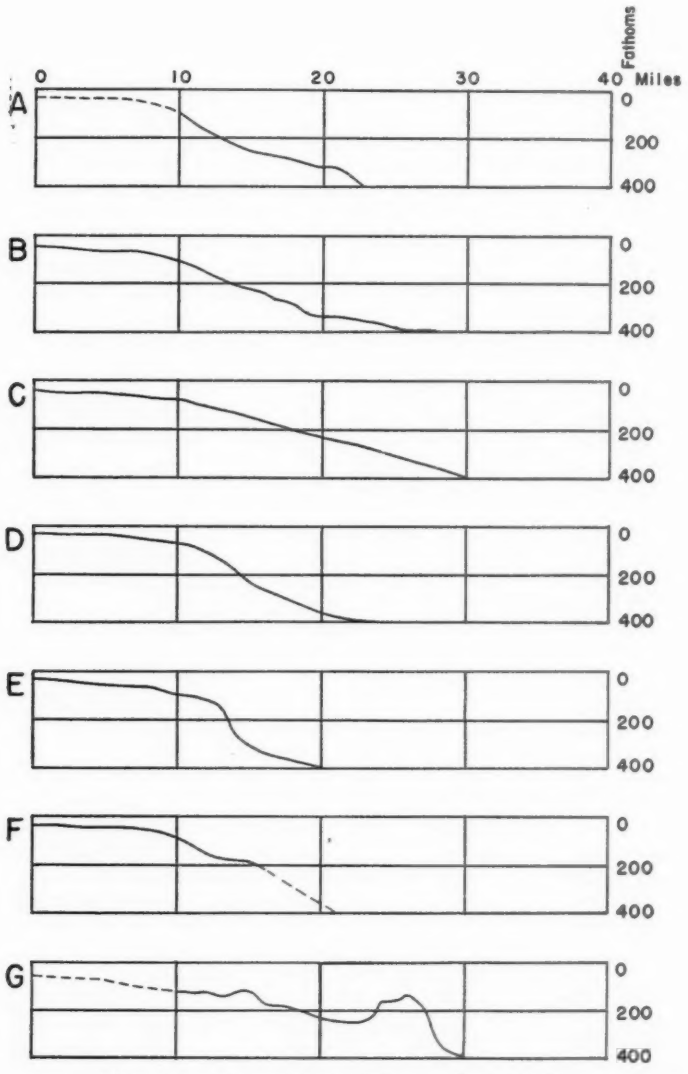


FIGURE 2.—PROFILES ACROSS BREAK IN SLOPE (A–G)

Locations shown on Figure 1. Solid portions of profiles taken from fathograms. Dotted portions taken from Plates 1 and 2. Sources of fathograms given in Appendix I. Vertical exaggeration × 17.

Gulf of Mexico described by Dietz and Menard (1951) and Dietz (1952, p. 1815) is evident on few of these profiles, even at the great vertical exaggeration of × 17.

If the break in slope is assumed to be a line along which the gradient increases to 10 fathoms per mile, the line is variable in depth. Off the mouth of the Rio Grande south of 26°40′ N. Lat. and in the vicinity of the Mississippi Trough, the line is at a depth of about 50 fathoms. Throughout most of the rest of the contoured area, it is near the 75-fathom curve

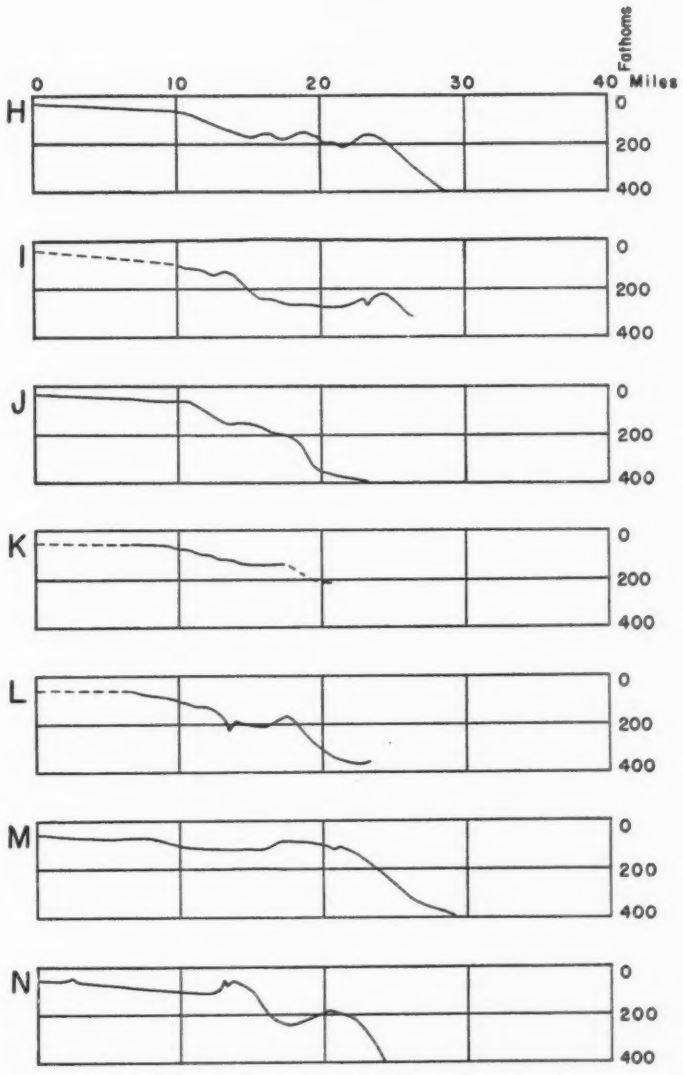


FIGURE 3.—PROFILES ACROSS BREAK IN SLOPE (H–N)

Locations shown on Figure 1. Solid portions of profiles taken from fathograms. Dotted portions taken from Plates 1 and 2. Sources of fathograms given in Appendix I. Vertical exaggeration × 17.

although locally it is deeper than 100 fathoms.

Many bioherms, described by Stetson (1953), rise to within a few fathoms of the surface near the break in slope between 90°00′ W.Long. and 95°00′ W.Long. The flat tops of the bioherms are covered with living corals and calcareous algae, and their steep flanks are covered with calcareous debris which grades into the surrounding fine-grained clastics. Many of these bioherms can be seen on Plates 1 and 2, and more would be indicated if a smaller contour interval had been used.

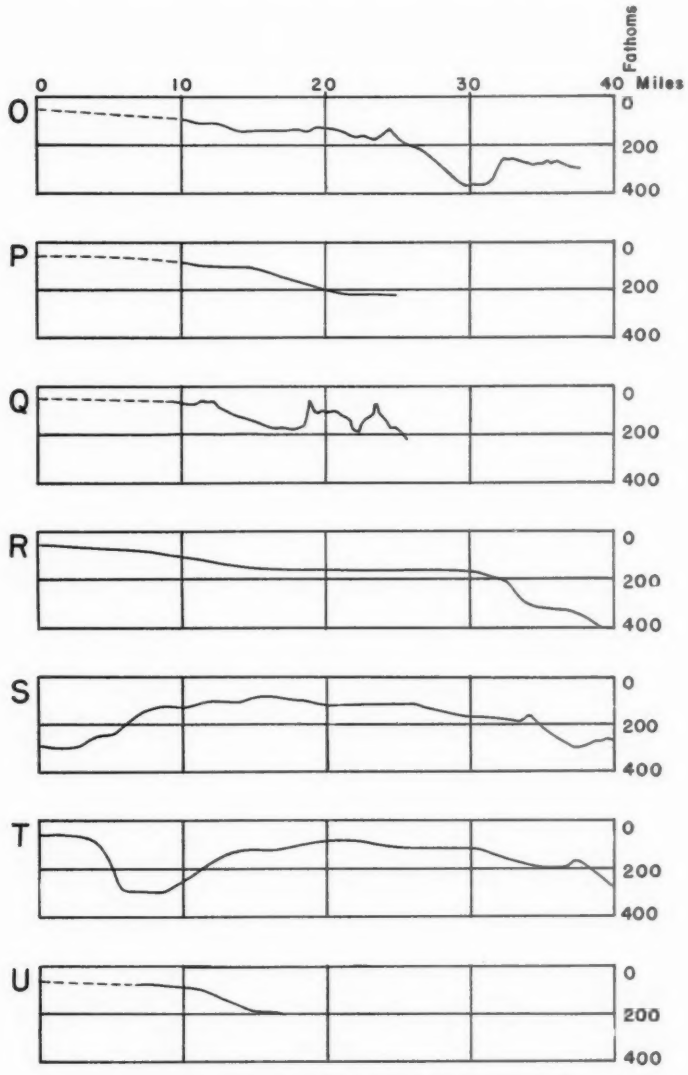


FIGURE 4.—PROFILES ACROSS BREAK IN SLOPE (O–U)
Locations shown on Figure 1. Solid portions of profiles taken from fathograms. Dotted portions taken from Plates 1 and 2. Sources of fathograms given in Appendix I. Vertical exaggeration × 17.

Sigsbee Deep

The Sigsbee Deep (Fig. 1) is a basin in the western Gulf of Mexico which extends from the base of the continental slope at about 1700 fathoms to a maximum depth of about 2200 fathoms. The basin has not been surveyed in detail, but available information indicates that slopes are gentle and relief subdued.

Continental Slope

General.—The continental slope (Fig. 1) in the northwest Gulf or Mexico extends from the edge of the continental shelf to the edge of the Sigsbee Deep. Between the 75- and 1700-fathom curves the slope is about 70 miles wide off the mouth of the Rio Grande, 135 miles wide at 92°00′ W.Long., and 70 miles wide west of the Mississippi Trough. The average gradient of the slope ranges between about 12 and 25 fathoms per mile, but these figures are not significant because of the varied topography of high relief on the continental slope. Locally, gradients exceed 500 fathoms per mile. The continental slope can be divided into four parts: (1) the Sigsbee Scarp, (2) the Trough Zone, (3) the Hummocky Zone, and (4) the Mississippi Trough.

Sigsbee Scarp.—An abrupt scarp, which the writer has termed the Sigsbee Scarp, is at the base of the continental slope (Fig. 1). The curves on the scarp were made from echo-sounding records taken by the Research Vessel ATLANTIS and surveying vessels of the Hydrographic Office of the U. S. Navy (Fig. 5). As positional control was relatively poor, the location is only approximate, and it can be assumed that topographic details are inaccurate. For instance, a greater concentration of lines of soundings permits more topographic detail to be shown northeast of 90°30′ W.Long. than between 93°00′ W.Long. and 94°00′ W.Long., although there is no evidence indicating smoother topography in either area. Also, it can be assumed that the scarp is steeper than shown.

The scarp is shown on Plate 1 between 95°10′ W.Long. and 94°10′ W.Long. at the lower edge of the contoured area; it is incised by a submarine trough at 94°30′. The scarp is shown also on Plate 2 trending northeast from the lower edge of the contoured area at 90°30′ W.Long. until it disappears at its junction with the Mississippi Trough.

The Sigsbee Scarp is comparable in size and relief to the fault scarp on the west side of the Wasatch Range, Utah, although the Wasatch Scarp is only about half as long as the Sigsbee Scarp.[2] That part of the Sigsbee Scarp shown on Figure 1 is more than 340 miles long, and it extends an unknown distance south off the east coast of Mexico. The scarp trends east from 95°10′ W.Long. to 91°40′ W.Long. where it turns abruptly northeast. It is 4800 feet high and has an average gradient of at least 200 fathoms per mile (1:5) at its western end. It decreases both in height and steepness to the east. The scarp is incised in places by troughs.

Another abrupt scarp about 350 miles long and trending roughly northwest occurs in the east Gulf of Mexico off the coast of Florida. The northwestern 150 miles of the scarp has been mapped and described by Jordan (1951); he found that it was about 7200 feet high and had an average gradient of about 500 fathoms per mile at 27°30′ N.Lat. and 85°30′ W.Long. The northwesternmost part forms the east flank of the lower part of the DeSoto Canyon, 125 miles east of the Mississippi Trough and beyond the limits of Figure 1. Between the DeSoto Canyon and the Mississippi Trough the continental slope trends northeast paralleling the strike of the easternmost part of the Sigsbee Scarp. The continental slope in this area has gentler gradients than those found on the Sigsbee Scarp, but they are steeper than the average gradient of the continental slope in the northwestern Gulf. The steepest slope in the area is about 60 fathoms per mile (1:17). The area is not considered part of the Sigsbee Scarp although it may be an extension of the same structure. Near the lower part of the DeSoto Canyon, the Florida Scarp turns north and intersects this northeast extension of the Sigsbee Scarp trend at about 60°.

Trough zone.—Above the Sigsbee Scarp a zone about 50 miles wide extends the length of the continental slope (shown in part on Pls. 1, 2). The zone consists of a surface of relatively low relief, typically illustrated in the area about

[2] Information concerning the Wasatch Range was taken from U. S. Geological Survey topographic sheets of the Timpanagos Cave, Orem, Aspen Grove, and Bridal Veil Falls quadrangles, Utah. The Wasatch Range is about 150 miles long and 7000 feet high from the base of the range at 5000 feet above sea level to the highest peaks at about 12,000 feet high. Average gradients on the scarp are as high as 3200 feet per mile but average less than 2300 feet per mile.

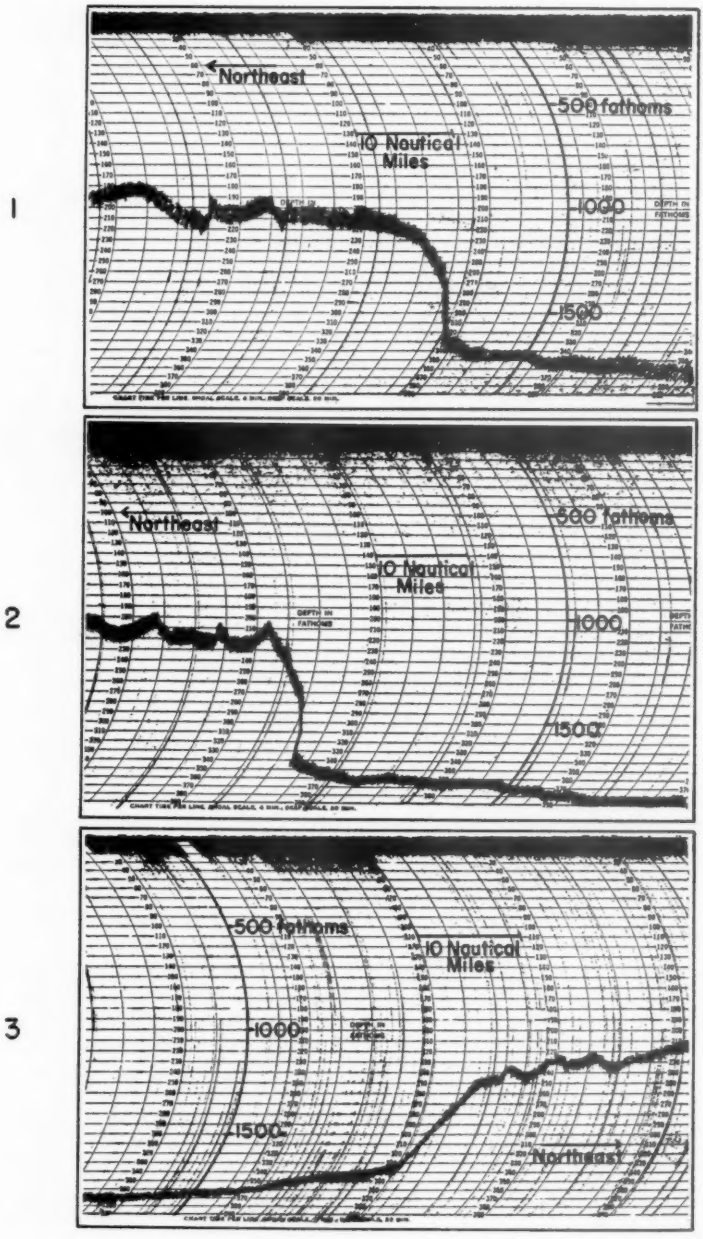

1
Northeast
500 fathoms
10 Nautical Miles
1000
1500
2
Northeast
500 fathoms
10 Nautical Miles
1000
1500
3
500 fathoms
10 Nautical Miles
1000
1500
Northeast

FIGURE 5 (1–3)

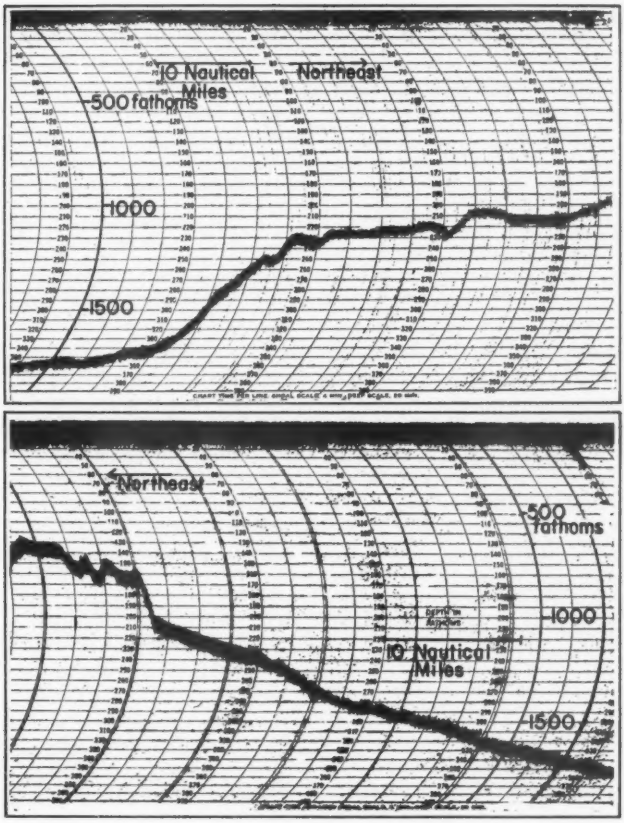


FIGURE 5 (4–5)

FIGURE 5.—FATHOGRAMS ACROSS SIGSBEE SCARP

Locations of profiles shown on Figure 1. All fathograms taken by surveying vessels of Hydrographic Office, U. S. Navy. (Vertical exaggeration = × 20 approximately)

(1) March 3, 1949; Book 91, Roll 37. Approximate position of base of scarp: 26° 11′ N. Lat., 93° 20′ W. Long.

(2) March 3, 1949; Book 85, Roll 30. Approximate position of base of scarp: 26° 01′ N. Lat., 93° 08′ W. Long.

(3) February 15–16, 1949; Book 61–62, Roll 24. Approximate position of base of scarp: 25° 53′ N. Lat., 92° 44′ W. Long.

(4) November 23, 1947; Book 53, Roll 30. Approximate position of base of scarp: 25° 45′ N. Lat., 92° 37′ W. Long.

(5) April 3–4, 1947; Book 16, Roll 20. Approximate position of base of scarp: 27° 27′ N. Lat., 89° 38′ W. Long.

95°10′ W.Long., 26°40′ N.Lat., and into which are cut many deep steep-sided troughs. These troughs extend down the slope and into the Sigsbee Deep and cut across the Sigsbee Scarp.

One complete trough is shown on Plate 2 at 94°20′ W.Long. In March 1951 a resurvey by the Research Vessel ATLANTIS corroborated the existence of the trough although additional data indicate that the location and shape indicated here are somewhat inaccurate. It is evident, however, that the trough is about 80 miles long, 3000 feet deep, and about 5 miles wide from rim to rim. The walls have average gradients of at least 3000 feet per mile (1:2), and locally some of the slopes may be nearly vertical. The trough is comparable to Yosemite Valley, California, in both size and shape.[3] The heads of other troughs are shown at the southern edge of the contoured area east of this trough (Pls. 1, 2). Fathograms taken by the Research Vessel ATLANTIS and surveying vessels of the Hydrographic Office show a number of troughs in the unsurveyed area north of the Sigsbee Scarp. Presumably these troughs join the heads shown on Plates 1 and 2. More closely spaced soundings with better positional control are necessary to determine their exact shape, size, and location. However, available information suggests that some are more than 3000 feet deep and that the slopes of their sides may exceed 3000 feet per mile (1:2). Some of the troughs have nearly flat bottoms such as the one at 91°00′ W.Long. at the southern edge of the contoured area on Plate 2; some are V-shaped such as the one that trends north-northwest from the southern edge of the contoured area at 92°20′ W.Long. (Pl. 2). From available soundings, many troughs cannot be contoured without depressions, the deepest being 1500 feet. The presence of depressions in submarine troughs is not uncommon, but proof of the existence of any of this magnitude must await further data.

Hummocky zone.—Between the break in slope at about the 75-fathom curve and the zone with troughs is a hummocky zone dominated by small hills and closed depressions, many of which are in areas of closely spaced soundings and excellent positional control. Although some troughs extend into this zone, they are not so plentiful or so striking. Two types of topographic patterns may be present in this zone:

Several topographic alignments are shown on Plate 3, and profiles across them are shown on Figure 7. Some depressions appear to be aligned with the troughs below them. Possibly the troughs formerly extended into the upper part of the slope and have been filled partially and irregularly by masses of sediment.

Mississippi Trough.—The Mississippi Trough (Pl. 2) cuts into the continental shelf and extends across the other three zones of the slope. It is about 20 miles wide and 1800 feet deep where it cuts across the break in slope at the edge of the continental shelf. The trough widens considerably in deeper water.

ORIGIN OF TOPOGRAPHY ON CONTINENTAL SLOPE

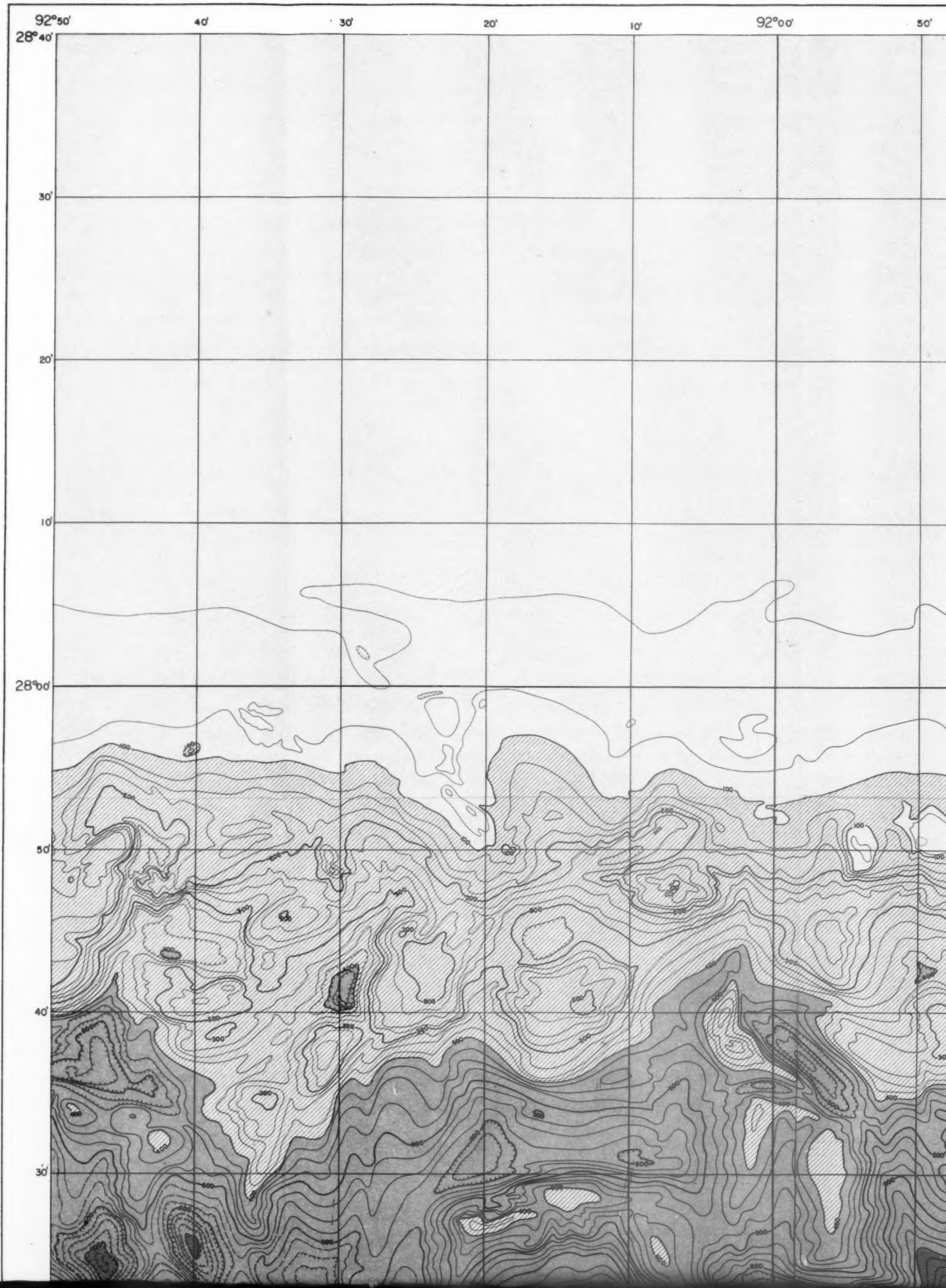
General

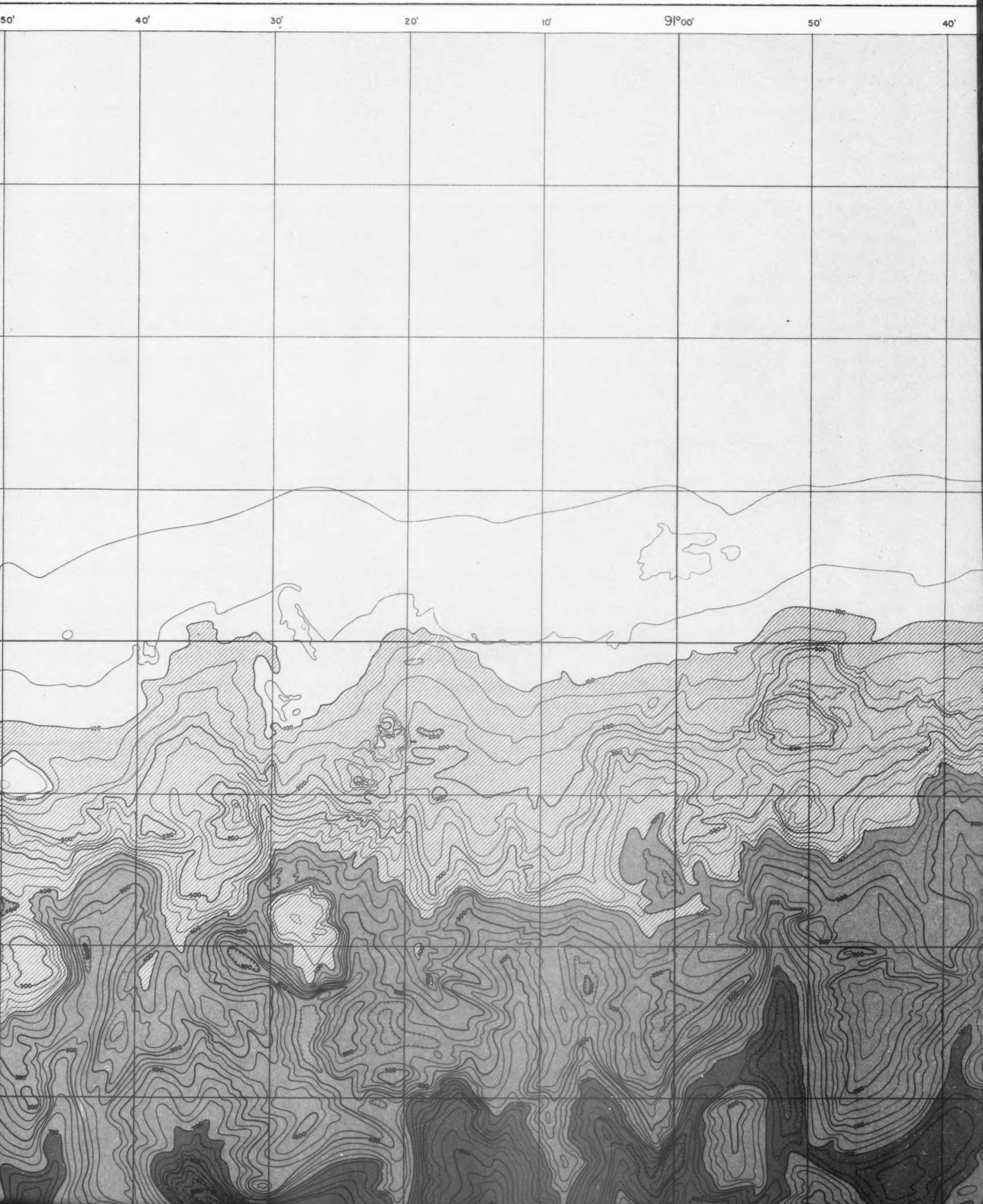
The topography on the continental slope of the northwest Gulf of Mexico may have been developed by crustal faulting, salt domes, subaerial erosion, and the failure of sediment under the stress of gravity.

Crustal Faulting

The Sigsbee Deep is bounded by scarps of such great extent and relief that only large crustal faults can explain them. The term "crustal fault" is used in contrast to smaller faults in sediments on the continental slope. The Florida Scarp off the west coast of Florida is about 350 miles long and at least 7200 feet high (Jordan, 1951). The Florida Straits are bounded by abrupt scarps probably formed by large normal faults (Pressler, 1947). The north and west sides of the Campeche Bank are bounded by scarps up to 8000 feet high (U. S. Coast and Geodetic Survey Chart No. 1007, Gulf of Mexico, 1927). The Sigsbee Scarp may represent a fault downthrown toward the Sigsbee Deep. In the northwest Gulf of Mexico the height of the continental slope may be in-

[3] Information concerning Yosemite Valley was taken from the U. S. Geological Survey Topographic map of Yosemite, California, 1938 edition. Yosemite Valley is about 2 miles wide from rim to rim and 2500 to 3500 feet deep. The valley floor is about three quarters of a mile wide. The walls of the valley are locally vertical, but average gradients range from 3000 to 8000 feet per mile.


92°50'
40'
30'
20'
10'
92°00'
50'
28°40'
30'
20'
10'
28°00'
50'
40'
30'


50'
40'
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91°00'
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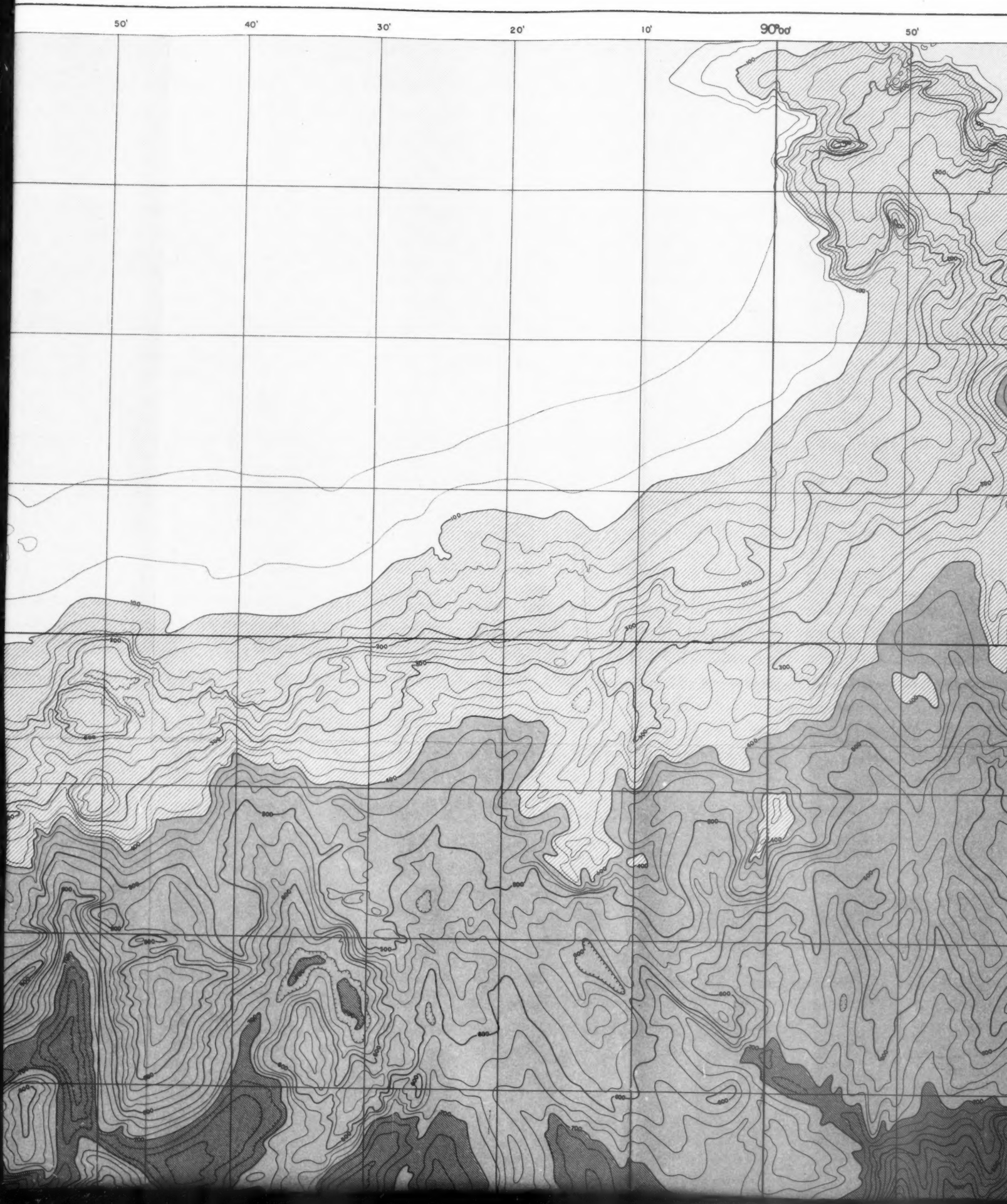
40'


50'
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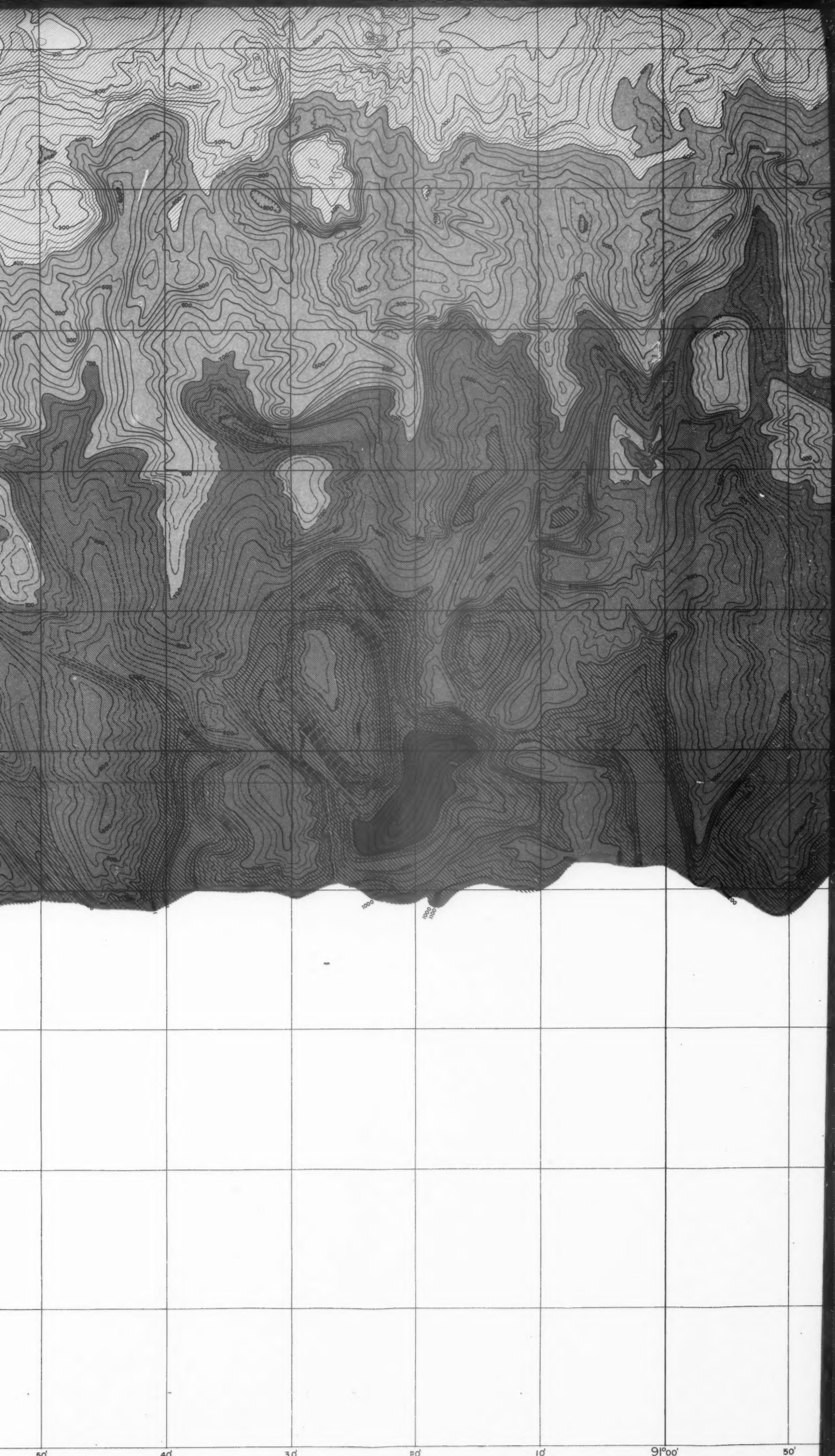
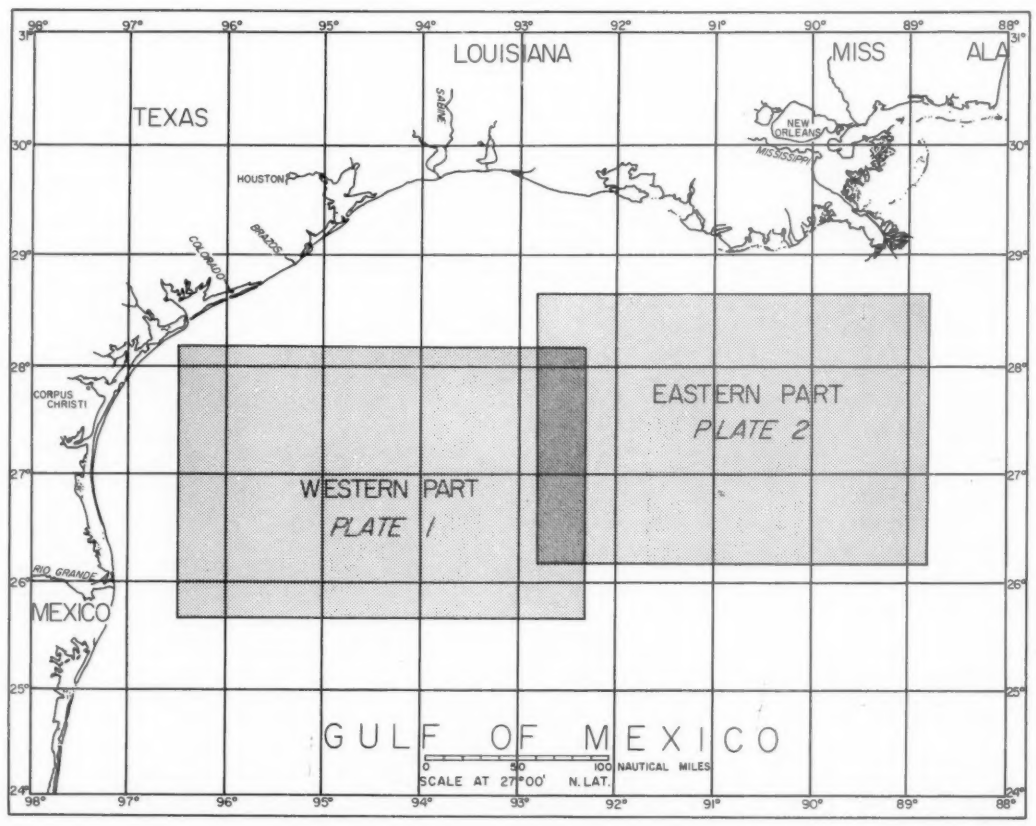
10'
92°00'

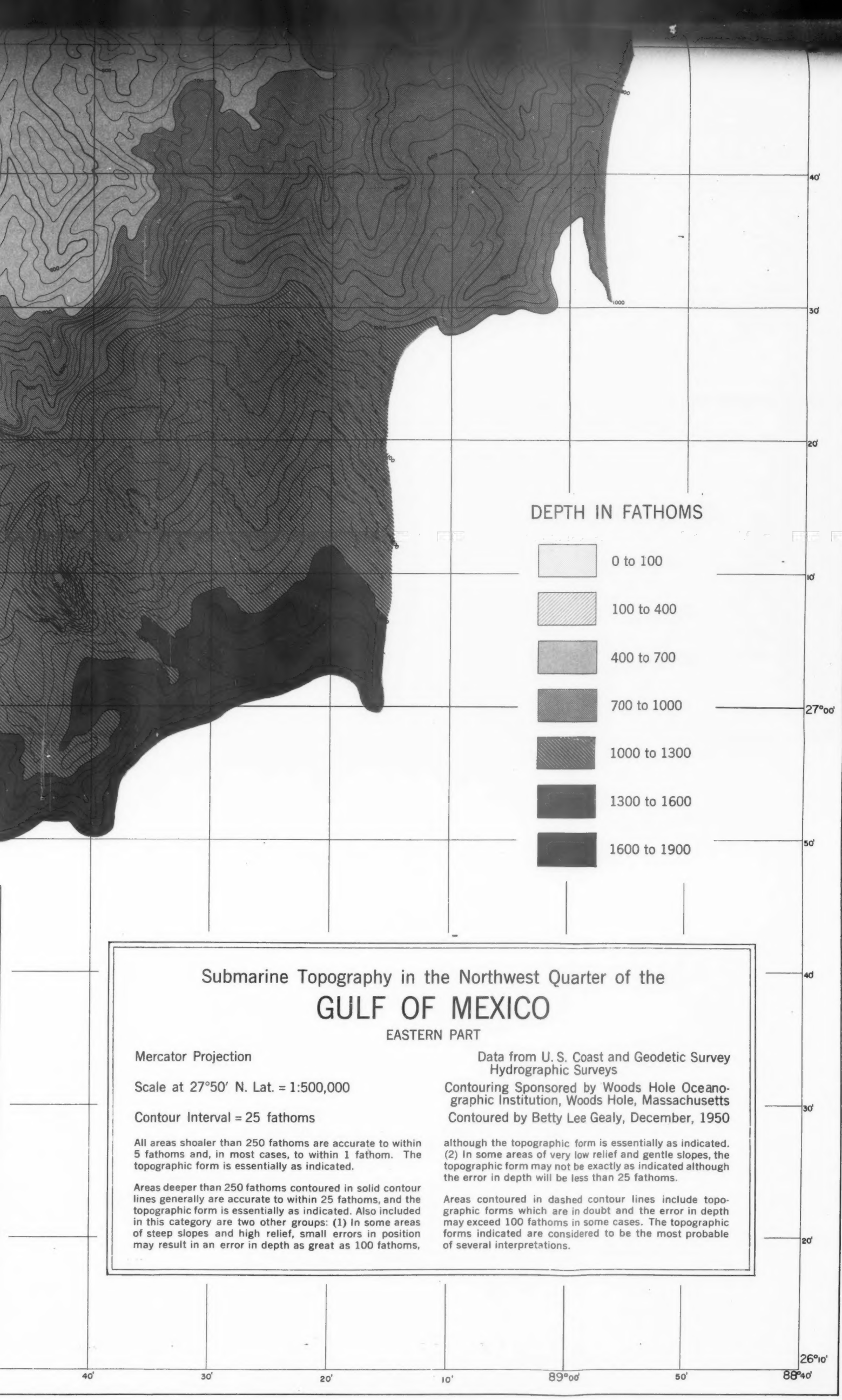


SUBMARINE TOPOGRAPHY IN THE NORTHWEST



91°00' 50' 40' 30' 20' 10' 90°00' 50'

IN THE NORTHWEST QUARTER OF THE GULF OF MEXICO EASTERN PART



Submarine Topography in the Northwest Quarter of the GULF OF MEXICO

EASTERN PART

Mercator Projection

Scale at 27°50′ N. Lat. = 1:500,000

Contour Interval = 25 fathoms

Data from U. S. Coast and Geodetic Survey Hydrographic Surveys

Contouring Sponsored by Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

Contoured by Betty Lee Gealy, December, 1950

All areas shoaler than 250 fathoms are accurate to within 5 fathoms and, in most cases, to within 1 fathom. The topographic form is essentially as indicated.

Areas deeper than 250 fathoms contoured in solid contour lines generally are accurate to within 25 fathoms, and the topographic form is essentially as indicated. Also included in this category are two other groups: (1) In some areas of steep slopes and high relief, small errors in position may result in an error in depth as great as 100 fathoms, although the topographic form is essentially as indicated. (2) In some areas of very low relief and gentle slopes, the topographic form may not be exactly as indicated although the error in depth will be less than 25 fathoms.

Areas contoured in dashed contour lines include topographic forms which are in doubt and the error in depth may exceed 100 fathoms in some cases. The topographic forms indicated are considered to be the most probable of several interpretations.

creasing owing to continued movement of this fault.

As the height of the Sigsbee Scarp is greater toward its western end, the throw of the fault may diminish progressively eastward. The fault may merge with a flexure between the Mississippi Trough (Fig. 1) and the DeSoto Canyon, 125 miles east of the Trough. However, because of the proximity of the Mississippi River, sedimentation may be more rapid on the eastern end of the scarp and the scarp partially masked by sediment. Two small troughs that cross the Sigsbee Scarp at 89°50′ W.Long. and 90°25′ W.Long. resemble hanging valleys. If they are hanging valleys, at least part of the faulting may postdate the cutting of the troughs. Otherwise, they may have been cut by an agency that does not necessarily produce accordant junctions.

The lower part of the DeSoto Canyon appears to have been formed by the junction of two major structural trends: the steep continental slope northeast of the Mississippi Trough, and the northern end of the Florida Scarp.

Salt Domes

There are a number of topographic features of very low relief on the continental shelf (R. A. Geyer, personal communication) which may indicate underlying salt domes. The features are roughly oval or circular and have diameters of a few miles to less than a mile. Also, the bioherms at the edge of the continental shelf may overlie other domes.

On the continental slope salt domes may have influenced some of the topography but there is no evidence that they have been an important geomorphic factor. The great size, high relief, and topographic form of most of the major features indicate that they were formed primarily by faulting, the failure of sediment under the stress of gravity, and, possibly, subaerial erosion.

Subaerial Erosion

Topographic evidence of subaerial erosion.—To test the theory of the subaerial erosion of the topography on the continental slope of the northwest Gulf of Mexico, Plate 3 was constructed showing "streams" in the valley bottoms and depressions filled with "lakes." The depths of the depressions from the sill depth to the deepest point are shown to the nearest 25 fathoms. This "stream-lake" pattern is used only as an aid in analyzing the topography and it is not presumed that streams and lakes ever existed on this surface as indicated.

Throughout most of the region troughs can be joined and tend to merge downhill as do subaerial valleys. In some areas, such as west of 94°30′ W. Long., the stream pattern is dendritic and resembles some subaerial stream erosion patterns. In other areas, such as between 94°00′ W. Long. and 93°00′ W. Long., the pattern suggests structural control.

Major streams shown by the technique have gradients of tens or hundreds of feet per mile. Most subaerial valleys of comparable size have gradients of a few inches or feet per mile. If these submarine troughs are the result of subaerial erosion their gradients must have been increased by: (1) a rate of sedimentation greater on the upper part of the continental slope than on the lower part, (2) tilting of the whole continental slope toward the Sigsbee Deep, (3) faulting, landslides, and folding producing local high gradients, and/or (4) submarine erosion.

Some of the submarine troughs contain depressions 250 fathoms deep, considerably deeper than depressions in most subaerial stream valleys. However, these submarine depressions can be explained by the unequal filling of pre-existing troughs or by large-scale faulting or folding.

Effects of Pleistocene lowering of sea level.—Authorities estimate that, during the maximum extent of glaciers during the Pleistocene, sea level was lowered from about 350 to 400 feet less an uncalculated amount attributable to isostatic adjustment (Flint, 1947, p. 437). Fisk (1944) has concluded, largely on the basis of the buried trenches of Pleistocene rivers on the Gulf Coastal Plain, that sea level was lowered about 450 feet. If sea level were lowered 450 feet, the shore line would have been near the present 75-fathom curve.

That Pleistocene rivers cut valleys on the continental shelf is implied by the fact that all the rivers entering the northwest Gulf of Mexico except the Rio Grande and Mississippi have estuaries, and by the fact that an entrenched Pleistocene Mississippi valley system is buried beneath Recent sediments (Fisk, 1944). However, the Mississippi Trough is the only topo-

graphic evidence of such valleys. Fisk (1944) finds that the bottom of the Late Wisconsin Mississippi Valley is between 400 and 450 feet below sea level in areas that show no evidence of recent warping or faulting. The valley is filled with sediment; the lower layers are gravel, and the upper levels finer material. Seismic surveys show that this valley continues offshore and trends in the direction of the Mississippi Trough (Carsey, 1950, p. 372).

The obvious conclusion is that the Mississippi Trough was cut by subaerial erosion during the Pleistocene when sea level was lowered. However, the present Mississippi Trough cuts the edge of the continental shelf to a depth of about 1800 feet, and it can be traced across the continental slope to at least the 1000-fathom curve. If sea level were lowered only 450 feet, the depth and extent of the trough implies either that the trough and buried valley system were eroded during lowered sea level of the Pleistocene or earlier and that the slope has been downwarped subsequently, or that at least that part of the Trough deeper than 75 fathoms was cut by submarine erosion.

Some indentations in the 100-fathom contour, such as those at 93°30′ W.Long., 92°18′ W. Long., and 91°35′ W.Long., may be related to other river valleys which crossed the continental shelf during lowered sea level of the Pleistocene but which subsequently have been filled with sediment.

During Pleistocene lowering of sea level great quantities of sediment would have been deposited near the edge of the continental shelf from: (1) material eroded from the continental shelf, (2) material removed when Pleistocene river valleys on the Gulf Coastal Plain were deepened, and (3) material normally transported by those rivers. Deposits of this material may be thick on the upper part of the continental slope, and the hummocky zone may be the surface of those deposits. These sediments should be thinner on the lower part of the continental slope and in the Sigsbee Deep. They should consist of weak unconsolidated and semiconsolidated material which would tend to move down the continental slope. Also, the mass should increase the gradient of the whole continental slope and make the slope less stable.

Submergence by downwarping.—The northwest Gulf of Mexico and the adjacent Coastal Plain are tectonically unstable. Nettleton (1952) reports about 40,000 feet of upper Jurassic and post-Jurassic sediments at the coast line. Wells drilled near the edge of the continental shelf have penetrated 16,000+ feet of upper Cenozoic sediment with shallow-water characteristics (Fisk, 1953). Near the Mississippi delta, Quaternary depoits are about 4000 feet thick, and post-Wisconsin sediments are more than 600 feet thick (Fisk, 1944). This indicates subsidence throughout the Cenozoic.

Geophysical and geological data are too meager to indicate the location, nature, or even the existence of the southern limb of the proposed Gulf Coast geosyncline. Well control in all cases is limited to on-shore and near-shore areas and to depths of less than 20,000 feet. Fisk (1953) has presented subsurface information indicating seaward thinning of several major stratigraphic units; he concludes that the axis of a geosyncline lies south of Houston, Texas, and near New Orleans, Louisiana. He claims that gravity relationships support this. Nettleton (1952) concludes from geophysical evidence that the regional dip of the basement underlying upper Jurassic and post-Jurassic sediments extends to the coast line without any indications of a synclinal axis at or immediately beyond the coast. At present it is not conclusive that the rate of post-Jurassic downwarp has been either greater or less under the continental slope than under the shore line.

Available evidence does not preclude that the continental slope could have subsided from above sea level to its present depth during the Cenozoic; much of the subsidence could have taken place during the Quaternary. The rate of subsidence could have been greater on the lower part of the continental slope, thus causing tilting toward the Sigsbee Deep. Irregular subsidence could have contributed to the irregularity of the topography on the continental slope.

Failure of Sediment under Stress of Gravity

General.—Under the northwest Gulf of Mexico a large, thick mass of unstable sediment is under shearing stress. There is evidence in the topography and sediments that the con-

tinental slope in this region is being modified by the failure of these sediments under stress.

The discussion of the behavior of sediments is based largely on literature in soil mechanics (Baver, 1948; Terzaghi and Peck, 1948; Knight, 1948). In addition, S. D. Wilson of the Division of Applied Science, Harvard University, has offered many suggestions.

Throughout the discussion the terms sand, silt, and clay refer only to size. The Wentworth grade scale is used where the term sand includes sediment with median diameters between 1 and $\frac{1}{16}$ mm, silt between $\frac{1}{16}$ and $\frac{1}{256}$ mm, and clay less than $\frac{1}{256}$ mm. "Clay minerals" refer to members of the illite, montmorillonite, kaolinite, and other hydrous silicate groups which are abundant in most clays and do not refer to grain size. This usage contrasts with that in the field of soil mechanics where all sediments consisting largely of clay minerals and possessing plasticity would be termed clay regardless of their grain size. Sediments with median diameters of clay would be called silt if they did not exhibit such plasticity, as their behavior under stress would correspond to that of silt.

Sources of stress.—Shearing stresses on the sediments under the northwest Gulf of Mexico result from several conditions: (1) An accumulation of sediment of uniform thickness on the continental shelf and slope, and on the floor of the Sigsbee Deep will tend both to increase the shearing stresses on the sediment and to consolidate and increase the shearing resistance of the underlying sediment. If consolidation does not take place before the shearing resistance is exceeded by the shearing stresses, the material will fail. (2) Sedimentation is probably not uniform over the whole area but is greater on the continental shelf and the upper part of the slope. This will increase the gradient of the continental slope and increase shearing stresses on the underlying sediment even more than an accumulation of uniform thickness. (3) The height of the continental slope is probably being increased by continued movement along a fault represented by the Sigsbee Scarp. As the height is increased, shearing stresses on the underlying material are increased. (4) Continued downwarping of the lower part of the continental slope may be increasing the angle of the slope. (5) While strong older sediments underlying the continental slope may maintain slopes of a given height and gradient, weak younger sediment being deposited on that slope may fail after reaching a critical thickness.

Sediments.—The sediments underlying the northwest Gulf of Mexico may be the largest mass of unstable sediment in the world. As the behavior of this sediment under shearing stress depends on the type of material involved, a summary of information concerning the sediments follows.

SURFACE AND NEAR-SURFACE SEDIMENTS: Recent surface and near-surface sediments, on the whole, are fine-grained, have low sorting coefficients and high water contents, and are unconsolidated. The upper 3 meters has been sampled during two cruises of the Research Vessel ATLANTIS of Woods Hole Oceanographic Institution. A reconnaissance survey of sedimentary conditions in the region was made in 1947 (Phleger, 1951a; Stetson, 1953; Trask, 1953). A critical area on the continental slope was surveyed in detail in 1951, and the data are being studied by the writer.

From 550 surface samples of sediment taken on the 1947 cruise of the ATLANTIS, Stetson (1953) has found that the continental shelf and slope in the northwest Gulf of Mexico is covered with silts, sands, and clays (Fig. 6). On the shelf, median diameters and the sorting coefficients of the sediments vary considerably. Sands as coarse as 0.1 mm are found in water as deep as 100 fathoms, and some clays are found in very shallow water. Below the 100-fathom curve the sediments are more uniformly distributed and consist of a fine, well-sorted clay (average median diameter: 0.0012 mm; average Trask sorting coefficient: 2.9).[4]

The 100 3-meter cores taken on the 1947 cruise of the ATLANTIS are being studied by H. C. Stetson, and the 15 3-meter cores taken on the 1951 cruise are being studied by the writer. Most of these cores can be correlated and show an upper zone of brown clay 5-60 cm thick with

[4] The Trask sorting coefficient is based on the ratio between the first and third quartiles: $So = \sqrt{Q_1/Q_3}$ where 75 per cent by weight of the sample consists of particles with diameters smaller than Q_1 and 25 per cent consists of particles smaller than Q_3 (Krumbein and Pettijohn, 1938, p. 230).

abundant globigerina. Beneath this clay is a mottled gray clay with some color banding. At a depth of 2 m the clay has a porosity greater than 30 per cent and as great as 70 per cent in some places. Mrs. Constance F. Klebba determined the water contents, and the writer converted them to approximate porosities. This sequence accumulated by normal particle sedimentation.

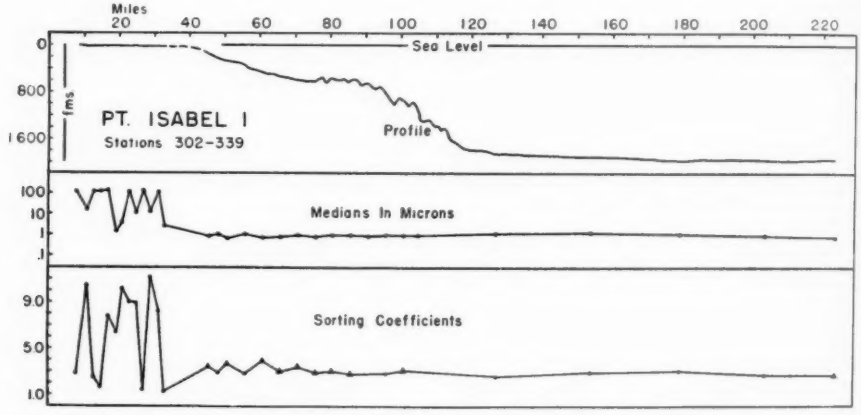


FIGURE 6.—SEDIMENTARY PROFILE ACROSS CONTINENTAL SHELF, CONTINENTAL SLOPE, AND INTO SIGSBEE DEEP

Traverse was made by ATLANTIS in 1951. Sedimentary analyses made by H. C. Stetson (1953). Location of profile shown on Figure 1. Vertical exaggeration on topographic profile × 17. (After H. C. Stetson)

SEDIMENTS AT DEPTH: There are no data concerning the sediments under the outer continental shelf, continental slope, and the Sigsbee Deep beneath the upper 3 meters, and the nature of these sediments must be postulated. Petroleum exploration programs have provided detailed information on Gulf Coastal Plain sediments to depths greater than 15,000 feet. Under the Gulf Coastal Plain, post-Paleocene sediments generally dip and thicken toward the present coast line; older beds dip more steeply than younger beds. The updip portions of the sediments consist largely of continental sandy and lignitic facies with some interbedded near-shore marine sands and clays (King, 1951, p. 169–174). Downdip these continental deposits interfinger with and grade into brackish-water and near-shore marine facies farther downdip. Near the coast, wells have not been drilled deep enough to penetrate Oligocene beds, but the upper Miocene and post-Miocene section consists largely of marine shales with minor amounts of sand. The sediments are highly consolidated but retain relatively high porosities. R. J. Granberry (Personal communication) says that Miocene montmorillonite shales in South Texas at depths greater than 17,000 feet have porosities as high as 25 per cent in places although most voids probably are filled with adsorbed water rather than free water. It is not uncommon to find sands with porosities of 25 per cent at this depth.

Under the outer continental shelf and slope sediments at depth probably resemble those found under the coast, but should be finer, on the whole, than their updip equivalents. However, at times of lowered sea level, as during the Pleistocene, coarse material may have been deposited far out into the Gulf. The sediments probably consist largely of clays and silts, with minor amounts of fine sand, and are highly consolidated. They should be much stronger than the near-surface sediments but probably are unstable under some conditions.

ANGLE OF REPOSE: The angle of repose of sediments underlying the continental slope in the northwest Gulf of Mexico depends on (1) the characteristics of the sediments, (2) the rate at which the slope increases, and (3) the mass of the sediments subject to failure.

The behavior of the whole mass of sediments

underlying the continental slope should be dominated by the behavior of silts and clays, for coarser sediments probably constitute only a small fraction. At depth the sediments may be highly consolidated and strong although they may retain relatively high porosities.

Consolidated fine sediment on the continental slope can form steeper slopes than unconsolidated or remolded sediment. If the overburden is removed from consolidated sediment by landslides or other agencies, this sediment will maintain steeper slopes than the adjacent unconsolidated surface material. However, removal of the overburden and exposure to water may cause the consolidated material to lose its original strength. Thixotropic effects will cause older unconsolidated sediment to have a higher angle of repose than younger material. Cross-bedding may increase the strength of a mass of sediment. However, failure may result from the inclusion of a single unstable layer in otherwise stable material.

If a slope is steepened slowly enough to permit complete drainage in the sediment, a steeper angle of repose can be acquired than if the slope is steepened suddenly and little drainage can occur. Thus, the rate of faulting or tilting will influence the angle of response of the sediments. A steeper angle of repose can be acquired if sediment accumulates slowly on the upper part of a slope than if it is deposited suddenly by a landslide.

Although vertical slopes a few feet high may be stable, slopes several hundred feet high in the same material may be unstable, because the total stresses tending to produce failure increase as the mass of the sediments increases.

Thus, the angle of repose is not constant for a given sediment but varies with the height of the slope and the rate at which the slope increases.

Types of failure.—When the angle of repose of the sediments underlying the continental slope is exceeded, the slope will fail by the downslope displacement of sediments. This displacement may be along a single shearing or fault plane, along several planes, or along so many planes that the failure would be termed plastic deformation or flow. The type of displacement depends on (1) the grain size, sorting coefficient, and mineralogy of the sediment; (2) the internal structure of the sediment resulting from its deposition, consolidation, and disturbance history; (3) the mass of the sediments involved; (4) the pressure under which the sediments are confined; and (5) the rate of stress application and the accompanying drainage conditions.

The terminology used for the types of failure in sediments depends on the presence of a well-defined shearing plane, the amount of mixing in the transported mass, the size of the failure, and the rate of movement. If a single shearing plane is present and if internal mixing in the transported mass is not great enough to destroy bedding and other internal structures, the failures commonly are called faults, if large, and landslides or slumps, if small. Confusion arises because fault is the term used for a surface of displacement, but landslide is the term used for the displaced material. If no single shearing plane is dominant and if the movement is relatively slow, the failure may be called creep and some features commonly called landslides or slumps are formed by creep. In either case, if the movement of the transported mass is dominated by flow characteristics and is relatively rapid, the failure may be called a sediment flow and would include mud flows and turbidity currents.

Divisions between these methods of transfer of sediment are arbitrary, and many failures may be intermediate ones. The amount of internal mixing is probably gradational between the stable transported masses of some faults and the most turbulent sediment flows. The transported mass may break into many large chunks which remain intact during movement but which are carried in a matrix of mixed sediment. A mass of sediment may move along an arcuate shearing plane at the upper end of the failure zone but may move by plastic failure or creep at the lower end of the failure zone. The transported mass may be so weakened during movement that, although the failure began as a fault, it may proceed as a sediment flow.

LANDSLIDES: Many smaller topographic features in the hummocky zone probably are due to landslides or small faults (Pl. 3, symbol no. 5). Downslope from some of the features are small hills which may represent the landslide deposits, indicating that the slides took place in relatively stable sediments. The absence of hills downslope from other features may imply that the landslide material, perhaps liquid, was transported a considerable distance down the slope.

These landslide features are confined to the hummocky zone on the upper part of the slope where relatively weak Pleistocene sediments should be thickest.

These features closely resemble the scars and deposits of many subaerial landslides. Landslides typically take place along narrow shearing zones or planes. The plane is usually curved, the radius of curvature being least at the upper end, greatest in the middle, and intermediate at the lower end. Commonly, a depression is formed at the upper end of the failure as the mass moves away from the slump scarp and down the slope under its own weight. The lower part fails passively owing to the thrust of the upper part, and the landslide deposit may be folded or faulted by thrusting. In layers of alternately weak and strong beds, failure is most likely along a weak stratum, and the shearing plane will not be a simple arc.

FAULTS: A number of topographic features on the continental slope may have been formed by faults (Pl. 3, symbol no. 3). Perhaps the most convincing feature is the scarp which trends due north from the southern edge of the map at 95°40′ W.Long. The scarp is over 3600 feet high, and a low area parallels the fault trace. The scarp is shown for over 20 miles, and it may extend beyond the southern limit of the map. The other topographic alingnments show no pronounced scarp, but only a steepened area on the upthrown side of the fault. The trace is marked by a line of low areas and depressions. Several profiles across these features are shown on Figure 7.

These features closely resemble structure-contour maps across some faults in Gulf Coastal Plain sediments. Faults are common throughout the western part of the Gulf Coastal Plain, and many of those not associated with salt domes have characteristic patterns. They tend to be concentric with the present coast and downthrown toward it (Kidd, 1939; King, 1951, p. 178). Beds on the downthrown side of these faults are bent down toward the fault plane in reverse drag, and a contour map of any stratigraphic horizon which crosses the fault shows a line of depressions, equivalent to sag ponds, next to the fault trace. In some cases, smaller antithetic faults parallel the major fault and are downthrown toward it on the downdip side, forming grabens (Cooper, 1937).

Many fault movements in the Gulf Coast were contemporaneous with deposition as some downthrown beds are thicker than their upthrown equivalents, and the two segments may consist of different facies (Kidd, 1939; Pool, 1940; Storm, 1945). Because of the contemporaneous deposition, the Gulf Coast faults may have an indistinct surface expression or disappear before reaching the upper beds, even though the maximum vertical displacement of a horizon at depth may amount to many hundreds of feet (Kidd, 1939; Pool, 1940). As faulting in the Gulf Coast apparently took place under conditions similar to those now existing in the northwest Gulf of Mexico, it is reasonable to expect faulting in the Gulf of Mexico.

Faults on the continental slope in the northwest Gulf of Mexico and many faults in the Gulf Coastal Plain may be the upper parts of large landslides. Geologists (Sheets, 1947; Storm, 1945) have noted the similarity between Gulf Coastal Plain faults and landslide features. In the Gulf Coastal Plain some fault planes decrease in dip with depth. In some cases, the vertical displacement diminishes not only upward, but also with depth as the dip of the fault plane decreases. These faults may be the upper parts of large landslides, the downthrown blocks moving along typical arcuate surfaces, steepest at their upper ends. The depth at which vertical displacement diminishes should be that depth at which movement of the block has a relatively large horizontal component. Only the upper part of the fault plane is commonly observed, and the middle and lower parts may not be recognizable from oil-well data. The failure plane may interesect the surface at its lower end, or the transported mass may be accomodated by plastic deformation.

CREEP: While creep may not produce effects recognizable in the topography, it may be of major importance in quantities of sediment transported over long periods of time. Creep is the movement of sediment down slopes by plastic failure where no single shearing plane is dominant. Sediments near the surface generally are least consolidated and least resistant to failure, but the stresses tending to produce failure increase with depth, so the rate of creep may not be greatest at the surface, as it is on subaerial slopes.

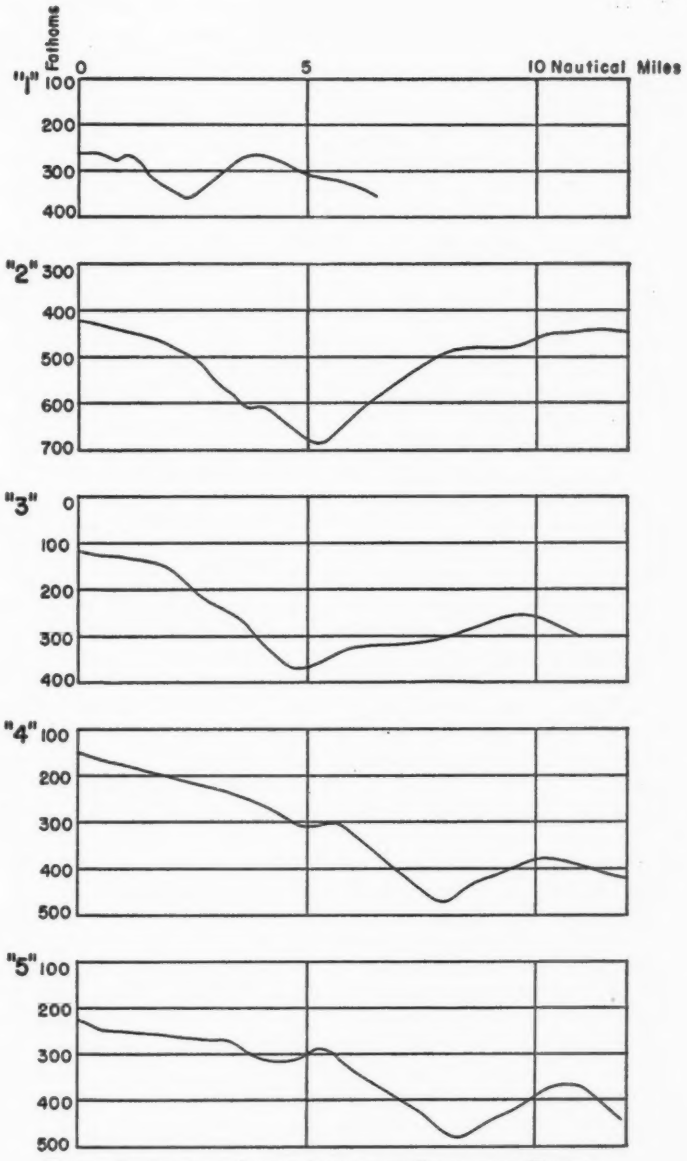


FIGURE 7.—PROFILES ACROSS ALIGNMENTS RESEMBLING FAULTS
Locations of profiles shown on Plate 3. Vertical exaggeration × 10.

SEDIMENT FLOWS: A sediment flow is a relatively rapid movement of sediment dominated by flow characteristics. It may be a viscous lamellar flow, a dilute turbulent flow, or any type between the two. A turbidity current is a sediment flow and may result from the failure of unstable sediments. Dietz (1952, p. 1807) suggests that the transition between a turbidity flow and a mud flow is where an increase in the density of the flow results in a decrease in

velocity. Kuenen (1950), in his flume experiments on turbidity flows, found that suspensions with densities as high as 2.2 (about 30 per cent porosity) could be made to flow down slopes. Denser suspensions moved by sliding. The change from sliding to turbulent flow was abrupt, and the density at which it occurred was a function of the slope, volume of the transported mass, and the grain-size distribution of the sediment. For convenience the term turbidity flow is confined to those sediment flows characterized by turbulence rather than to a change in density or a change in velocity, either of which may be more difficult to observe. Sediment flows may be made up largely of chunks of relatively intact sediment carried in a matrix of liquefied sediment. If a layer of unstable sediment is buried, the sediment may liquefy and act as a gliding plane. Thus, some landslides and faults may result from the liquefaction of sediment at depth.

Saturated loose sands, unconsolidated clays with high water contents, consolidated quick-clays with high water contents, and silts may be transformed into suspensions or semiliquids which can flow down gentle slopes for great distances. These sediments can flow without increasing the water content. However, subsequent dilution may take place after flow has begun, and the distance of the flow may be increased thereby.

In the hummocky zone a number of amphitheaterlike depressions with narrow outlets closely resemble bottleneck scars characteristic of flow failures. An excellent example of such a scar is one 35 meters deep left by the 1893 landslide in Vaerdalen, Norway. A mass of quick-clay over a large area was transformed into a semiliquid which flowed through a narrow outlet into the valley below (Holmsen, 1953).

The steep headwalls of many of the troughs in the northwest Gulf of Mexico suggest that the troughs may have been enlarged and lengthened by landslide-sediment flow mechanisms, and some of the troughs may have been formed completely by them. Landslides moving along arcuate surfaces leave relatively steep slump scarps. If the transported mass is removed from the area by flow the scarp remains unsupported and may be a site for subsequent landslides. Many of the troughs off the Atlantic coast of North America also have steep headwalls. The writer considers this landslide-sediment flow mechanism the most effective submarine erosion agency yet proposed for the formation of submarine troughs.

The topographic form of most of the submarine troughs in the northwest Gulf of Mexico indicates that they have been formed predominantly through erosion, but the topography does not indicate whether the erosion was subaerial or submarine. Once in motion, submarine sediment flows may erode the surface over which they move. That subaerial sediment flows can erode is well known. Crandell (1952) described what he termed an earth-flow in South Dakota which began as a landslide in the upper part of a valley, left a slump scarp, and moved 350 feet down the valley. The transported mass consisted of debris with a high sorting coeffecient and was 10–20 feet thick and 85 feet long. The moving mass gouged extensive slickensides along the walls of the gully, indicating that the flow actually eroded the surface over which it moved. Blackwelder (1928) describes a mud flow which moved down a canyon on the west side of the Wasatch Range, Utah, and left the walls of the canyon torn and scarred. Submarine sediment flows may move more slowly and be less effective agencies of erosion than either subaerial sediment flows or normal streams. However, that the troughs in the northwest Gulf of Mexico were eroded by submarine sediment flows remains a possibility.

Sediment flows are depositional agencies, but some beds deposited by them are difficult to distinguish from those deposited by normal particle sedimentation. Graded beds may be deposited under several conditions (Kuenen, 1950) including deposition by sediment flows. However, not all beds deposited by sediment flows will be graded. A graded bed requires that the sediment have a high sorting coefficient and that it be mixed, and that the flow have a viscosity low enough to permit differential settling. If the sediment has a low sorting coefficient and remains uncontaminated, grading may not be apparent in the resulting deposit. If the sediment has a high sorting coefficient, if it is thoroughly mixed during flow, and

if the viscosity of the flow is high, the resulting deposit should consist of a homogeneous mass of material with a high sorting coefficient.

The presence of a chronological sequence of layers is not proof that the sequence was not deposited by a sediment flow for sediments may flow without vertical mixing. In 1938 a large section of the upstream face of the Fort Peck Dam, Montana, and the underlying material moved downslope about 1400 feet in about 3 minutes. Because of the speed and distance of the movement and because the resulting deposit had a nearly level surface, some engineers believe that the material must have been liquid during movement and that movement was by flow (Middlebrooks, 1942; Casagrande, 1950). The stratigraphic sequence in the dam material and the underlying sediment was preserved during flow and was present in the resulting deposit, although the thicknesses and horizontal extent of the beds had been altered. The implication is that the flow was lamellar. Therefore, sections of layered sediment may move down slopes for great distances and still retain the original stratigraphic sequence.

In the northwest Gulf of Mexico, sediments believed to have been deposited by sediment flows have been found in two localities. The first is in the bottom of a steep-sided submarine trough where cores 229 and 232 were taken by the ATLANTIS in 1951 (Fig. 1). The cores were taken 7 miles apart, 135 miles from the coast, and at a depth of 1050 fathoms. The trough has a flat bottom in contrast to other V-shaped troughs in the region. Both cores consist of many layers from less than 1 mm to about 20 cm thick, but they cannot be correlated lithologically. The layers consist of sand, silt, and clay which contain Foraminifera, pteropod, and fish fragments. Sorting coefficients are high in some layers, low in others. Some of the layers contain clay balls. The sequences in these two cores are radically different from those in cores not taken in submarine troughs. The layers differ so much in grain size, sorting, and fossil content that, if they were not deposited by sediment flows, sedimentary conditions must have changed more radically and frequently than is believed possible in such deep water so far from the coast. It is unlikely that these are older sediments into which the trough has been cut because the sediments are as unconsolidated as other sediments known to be late Pleistocene-Recent by their Foraminifera content. The trough is about 3000 feet deep, and, even if the height of the walls has been increased by sedimentation while the bottom has been kept clear by sediment flows, the trough probably was cut into consolidated sediment. The U shape of the trough is considered to be due to partial filling of the trough by debris deposited by less viscous sediment flows. Similar layers of debris may be expected in other U-shaped troughs in the region, but not in the V-shaped troughs.

The second locality in which sediments believed to have been deposited by a sediment flow have been found is in the Sigsbee Deep. During the 1947 cruise of the ATLANTIS to the northwest Gulf of Mexico, cores 335 and 336 (Fig. 1) were taken 30 miles apart, 120 miles from the mouth of the Rio Grande, at a depth of 1900 fathoms. The cores have been described in part by Phleger (1951b) and consist of three layers: (1) an upper clay, (2) a middle sand, and (3) a lower clay. The upper clay is about 1.5 m thick, and Phleger (1951b, p. 71) reports benthonic Foraminifera which are predominantly deep-water types and planktonic Foraminifera characteristic of warm water. The Foraminifera and grain size of the upper clay are both characteristic of the locality and depth at which they are found. The middle sand is 32 cm thick in core 335 and 86 cm thick in core 336. The sand has a median diameter of 0.1 mm and a Trask sorting coefficient of 1.3. No graded bedding is apparent. Phleger (1951b, p. 72) says that the benthonic Foraminifera in the sand consist of a few deep-water types, some intermediate types, and many types characteristic of water shoaler than 50 fathoms. The planktonic Foraminifera are those characteristic of cold water; the sand is probably equivalent to the last glacial period. The lower clay was recovered in core 336 only.

Sands as coarse as this are found as surface sediment only in water shoaler than 100 fathoms and in most cases shoaler than 50 fathoms. Considering the grain size and Foraminifera content, the sand must either have been submerged to its present depth by downwarp of the ocean floor or by rising sea level, or have been transported to its present position. As a

post-Wisconsin submergence of 11,000 feet seems unlikely, transportation of the sand to its present site seems a preferable theory. Since sea level was probably near the present 75-fathom curve during late Pleistocene time the site of accumulation need not have been farther inland than the edge of the continental shelf. A surface deposit of loose sand near the edge of the shelf may have liquefied and flowed down the steep continental slope. Such a flow could have moved down the large submarine trough north of the present position of the sand.

Phleger (1951b, p. 74) says that the inclusion in the sand of fauna characteristic of intermediate and deep water in addition to shallow water must be attributed to one of three conditions: (1) the sand moved down the slope in stages, being repopulated with fauna characteristic of the depth at each resting place; (2) the supply of sand came from several sources and depths; or (3) the sediment was mixed during one movement. If Foraminifera characteristic of all depths were mixed with or added to the original sand, finer sediment characteristic of those depths must also have been mixed with or added to the original sand. The fact that the present sand has a sorting coefficient lower than most beach sand indicates that sorting by grain size took place during movement and that fractions finer than the sand were carried beyond cores 335 and 336.

Conclusions

Several hypotheses concerning the origin of the topography on the continental slope in the northwest Gulf of Mexico present themselves when the agencies which may be operating in the Gulf and the sediments underlying the region are considered:

(1) A number of features of low relief on the continental shelf and the bioherms at the edge of the continental shelf have topographic forms and dimensions suggesting they overlie salt domes. On the continental slope salt domes may have influenced some of the topography, but there is no evidence which indicates that they have been an important geomorphic factor.

(2) The Sigsbee Deep is bounded by scarps of such great extent and relief that large crustal faults are the only reasonable explanation of them. The Sigsbee Scarp probably represents a fault downthrown toward the Sigsbee Deep. The height and gradient of the scarp indicate that faulting is greatest at the western end of the scarp, diminishes near the Mississippi Trough, and merges into a flexure between the Mississippi Trough and the DeSoto Canyon. Hanging valleys indicate either that at least part of the faulting took place after the troughs on the continental slope were cut, or that the troughs were cut by submarine agencies which produce hanging valleys as a normal process of erosion. The northeast side of the Sigsbee Deep is shoaler than the northwest side, owing either to heavier sedimentation in the east or greater downfaulting in the west.

(3) In the northwest Gulf of Mexico a large mass of unstable sediments is under stress. Evidence in the topography and sediments indicates that this mass is failing by landslides, faults, creep, and sediment flows. Divisions among these types of failure are arbitrary, and many failures may be intermediate. Most of the topographic features on the continental slope probably have been formed completely or modified extensively by the failure of these sediments under the stress of gravity.

(4) The sand beneath a meter of clay in the Sigsbee Deep may have accumulated near the edge of the continental shelf during lowered sea level of the late Pleistocene and transported down the continental slope to its present position. As the sand contains benthonic Foraminifera characteristic of shallow, intermediate, and deep water, those forms must have been mixed with the sediment during movement. As the finer sediments characteristic of intermediate and deep water are not present in the sand, it is assumed the sediment was sorted during movement and the fine fractions carried beyond the sand.

(5) The topographic form of most of the submarine troughs in the northwest Gulf of Mexico indicates they were formed predominantly by erosion. The topography does not indicate conclusively what agencies were primarily responsible for that erosion. The writer believes the most reasonable hypothesis for the formation of these troughs is a composite one involving four phases:

(a) The original troughs were formed by

subaerial erosion prior to the formation of the Gulf of Mexico. Although the landslide-sediment flow mechanism is considered by the writer the most effective submarine erosion agency yet proposed for the formation of submarine troughs, she is hesitant to ascribe the formation of troughs 60 miles long and 3000 feet deep to this mechanism.

(b) After the troughs were formed the continental slope was submerged and tilted. The basement at the coast of the northwest Gulf of Mexico has been downwarped about 40,000 feet since the Jurassic, and the basement under the continental slope may have been downwarped an equal amount. Thick Pleistocene and Recent sediments under the coast line indicate that the basement is still subsiding. The continental slope could have subsided from above sea level to its present depth during the Cenozoic, and much of the subsidence could have taken place during the Quaternary.

(c) The upper parts of the troughs were masked by sediments deposited when sea level was lowered during the Pleistocene. The hummocky zone on the upper part of the continental slope represents the surface of these sediments. Few troughs extend into this zone; those present are more shallow, and their form is more irregular than those on the lower part at the slope.

(d) The troughs were modified extensively, largely by the failure of sediments under stress. The troughs may have been enlarged and lengthened after submergence by landslide-sediment flow mechanisms, and some troughs or parts of troughs may have been formed completely by them. The relatively unconsolidated Pleistocene sediments on the upper part of the slope should furnish an ample source of supply for the sediment flows. Although submarine sediment flows probably are less effective agencies of erosion than either subaerial sediment flows or normal streams, some part of the troughs could have been deepened by them.

The deep depressions in the troughs could have been formed during or after submergence by faults which cross the troughs, partial and irregular filling of the troughs by landslide debris and viscous sediment flows, or folding accompanying irregular subsidence.

Those troughs with flat bottoms have been filled partially with many thin layers of sediment, and it is presumed that this fill is absent in V-shaped troughs. The thin layers of sediment probably have been deposited by less viscous sediment flows.

Some aspects of this composite hypothesis for the origin of submarine troughs resembles a hypothesis presented by Shepard (1951a; 1951b; 1952).

(6) The upper part of the Mississippi Trough may be related to the Pleistocene Mississippi River Valley, including the buried off-shore extension of the Valley on the continental shelf. The upper part of the Mississippi Trough and a buried extension on the continental shelf could have been eroded subaerially during lowered sea level of the Pleistocene. If that part of the Mississippi Trough deeper than the 75-fathom curve were eroded subaerially, it must have been downwarped subsequently to its present depth. At least the lower part of the Mississippi Trough may be the same age as other troughs on the continental slope, and the upper part could have been re-opened by subaerial erosion during the Pleistocene.

Some of the indentations in the 100-fathom cruve may be related to other Pleistocene river valleys which crossed the continental shelf and which have been filled with sediment.

(7) On the upper part of the continental slope is a hummocky zone which probably represents the surface of thick Pleistocene sediments deposited when sea level was near the present 75-fathom curve. The presence of many topographic features which resemble landslide scars, landslide deposits, and sediment-flow scars indicates that this mass of sediments is relatively unconsolidated and weak and that the stresses tending to produce failure have increased rapidly. These relatively weak sediments are probably the source of many sediment flows which move down the troughs below the zone into the Sigsbee Deep.

The absence of a hummocky topography and the presence of relatively steep slopes on the lower part of the continental slope indicate that sediments underlying this region are more stable and less subject to failure than are those on the upper part. Pleistocene sediments may be thin on the lower part of the slope, and older

more consolidated sediments may be near the surface.

As only a few submarine troughs of modified form are found in this upper zone, probably the troughs are older than these Pleistocene sediments and the sediments have filled the upper parts of the troughs. Some depressions in the hummocky zone are aligned with troughs below them, indicating that the upper parts of the troughs have been partially and irregularly filled with sediment.

(8) The red lines shown on Plate 3 follow a series of depressions and low areas which may correspond to sagponds near fault traces. Above the lines are steep slopes which may be fault scarps. These faults are probably large arcuate shear planes along which large masses of sediment have failed. These alignments may indicate only a few of the many faults still active on the continental shelf and slope. Other faults may leave less pronounced topographic expressions or may be masked or obliterated by sedimentation.

Appendix I

Sources of fathograms used in constructing profiles across break in slope (Figs. 2, 3, 4). Locations of profiles shown on Figure 1.

"A"—Atlantis, 1947. Port Isabel I, stations 314–324
"B"—Atlantis, 1947. Port Isabel II, stations 276–286
"C"—Atlantis, 1947. Aransas, stations 256–269
"D"—Atlantis, 1947. Matagorda II, stations 169–179
"E"—Atlantis, 1947. Brazos I, stations 117–119
"F"—Atlantis, 1947. Brazos II, stations 66–74
"G"—Atlantis, 1951. March 7, 1300–1534
"H"—Atlantis, 1947. Galveston, stations 346–364
"I"—Atlantis, 1951. March 9, 0118–0318
"J"—Atlantis, 1947. Heald Bank, stations 25–37
"K"—Atlantis, 1951. March 8, 0537–0656
"L"—Atlantis, 1951. March 8, 0326–0537
"M"—Atlantis, 1947. Trinity Shoal I, stations 436–450
"N"—Atlantis, 1947. Trinity Shoal II, stations 485–492
"O"—Hydrographic Office A. March 11, 1949; Book 91, Roll 43, 0708–0947
"P"—Hydrographic Office E. February 16, 1949; Book 84, Roll 35, 1122–1300
"Q"—Hydrographic Office G. February 16, 1949; Book 62, Roll 24, 1200–1410
"R"—Atlantis, 1947. Ship Shoal, stations 525–534
"S"—Hydrographic Office K. January 5, 1949; Book 38, Roll 19, 1900–2400
"T"—Hydrographic Office O. February 11, 1947; Book 9, Roll 20, 0930–1400
"U"—Hydrographic Office N. March 7, 1949; Book 63, Roll 25, 2249–2310

References Cited

Baver, L. D., 1948, Soil physics: N. Y., John Wiley and Sons, 398 pages.

Blackwelder, Eliot, 1928, Mudflow as a geologic agent in semi-arid mountains: Geol. Soc. America Bull., v. 39, p. 465–484.

Carsey, J. Ben, 1950, Geology of Gulf Coastal area and Continental Shelf: Am. Assoc. Petroleum Geologists Bull., v. 34, p. 361–385.

Casagrande, Arthur, 1950, Notes on the design of earth dams: Boston Soc. Civil Eng. Jour., v. 37, p. 417–423.

Colle, Jack, Cookes, W. F., Jr., Denham, R. L., Ferguson, H. C., McGuirt, J. H., Reedy, Frank, Jr., and Weaver, Paul, 1952, Sedimentary volumes in Gulf Coastal Plain of United States and Mexico, part IV, volume of Mesozoic and Cenozoic sediments in Western Gulf Coastal Plain: Geol. Soc. America Bull., v. 63, p. 1193–1200.

Cooper, Herschel H., 1937, Occurrence and accumulation of oil in Laredo district, Texas: Am. Assoc. Petroleum Geologists Bull., v. 21, p. 1422–1438.

Crandell, Dwight R., 1952, Landslides and rapid-flowage phenomena near Pierre, South Dakota: Econ. Geology, v. 47, p. 548.

Dietz, Robert S., 1952, Geomorphic evolutions of Continental Slope terrace (Continental Shelf and Slope): Am. Assoc. Petroleum Geologists Bull., v. 36, p. 1802–1819.

Dietz, Robert S., and Menard, Henry W., 1951, Origin of abrupt change in slope at Continental Shelf margin: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 1994–2016.

Fisk, Harold N., 1944, Geological investigation of the alluvial valley of the lower Mississippi River: War Dept., Corps Engineers, U. S. Army, Miss. River Comm., Vicksburg, Miss., 78 pages.

—— 1953, Setting of Gulf Coast geosyncline: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 1129.

Flint, Richard Foster, 1947, Glacial geology and the Pleistocene epoch: N. Y., John Wiley and Sons, 589 pages.

Holmsen, Per, 1953, Landslips in Norwegian quick-clays: Inst. Civil Engineers, Geotechnique, v. III, p. 187–194.

Jordan, G. F., 1951, Continental Slope off Apalachicola River, Florida: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 1978–1993.

Kidd, Gentry, 1939, Developments in South Texas, 1938–1939: Am. Assoc. Petroleum Geologists Bull., v. 23, p. 860–870.

King, Philip B., 1951, The tectonics of middle North America: Princeton, N. J., Princeton Univ. Press, 203 pages.

Knight, B. H., 1948, Soil mechanics for civil engineers: London, E. Arnold and Co., 255 p.

Koppejan, A. W., Wamelen, B. M., and Weinberg, L. J. H., 1948, Coastal flow slides in the Dutch Province of Zeeland: Rotterdam, Proc. 2nd Internat. Conf. Soil Mech. and Found. Engineering, v. 5, p. 89–96.

Krumbein, W. C., and Pettijohn, F. J., 1938, Manual of sedimentary petrography: N. Y., Appleton-Century-Crofts, Inc., 549 pages.

Kuenen, Ph. H., and Migliorini, C. I., 1950, Turbid-

ity currents as a cause of graded bedding: Jour. Geology, v. 58, p. 91–127.

Middlebrooks, T. A., 1942, Fort Peck slide: Am. Soc. Civil Engineers Trans., Paper no. 2144, v. 68, p. 723–764. Discussions by Jacob Feld, Joel D. Justin, William Gerig, Alfred J. Ryan, Glennon Gilboy, Frank E. Falquist, Irving B. Crosby, and T. A. Middlebrooks.

Nettleton, L. L., 1952, Sedimentary volumes in Gulf Coastal Plain of United States and Mexico, part VI, Geophysical aspects: Geol. Soc. America Bull., v. 63, p. 1221–1228.

Phleger, Fred B, 1951a, Ecology of Foraminifera, Northwest Gulf of Mexico, part I: Foraminifera distribution: Geol. Soc. America Mem. 46, p. 1–88.

—— 1951b, Displaced Foraminifera faunas, in turbidity currents and the transportation of coarse sediments to deep water, a symposium: Soc. Econ. Paleontologists and Mineralogists, Special Pub. no. 2, p. 66–75.

Pool, J. C., 1940, Saxet oil and gas field, Nueces County, Texas: Am. Assoc. Petroleum Geologists Bull., v. 24, p. 1805–1835.

Pressler, E. D., 1947, Geology and occurrence of oil in Florida: Am. Assoc. Petroleum Geologists Bull., v. 31, p. 1851–1862.

Rosenqvist, I. Th., 1953, Considerations on the sensitivity of Norwegian quick-clays: Inst. Civil Engineers, Geotechnology, v. III, p. 195–200.

Sebring, Louie Jr., 1948, Slick-Wilcox field, DeWitt and Goliad Counties, Texas: Am. Assoc. Petroleum Geologists Bull., v. 32, p. 228–251.

Sheets, Martin, 1947, Diastrophism during historic time in Gulf Coastal Plain: Am. Assoc. Petroleum Geologists Bull., v. 31, p. 201–226.

Shepard, Francis (1951a), Submarine canyons: A joint product of rivers and submarine processes: Science, v. 114, no. 2949, p. 7–9.

—— (1951b), Submarine erosion, a discussion of recent papers: Geol. Soc. America Bull., v. 62, p. 1413–1418.

—— (1952), Composite origin of submarine canyons; Jour. Geology, v. 60, p. 84–96.

Stetson, Henry C., 1953, The continental terrace of the Western Gulf of Mexico: its surface sediments, origin and development: Mass. Inst. Technology and Woods Hole Oceanographic Inst., Papers in Physical Oceanography and Meteorology, v. 12, no. 4, pt. I, p. 1–45.

Storm, L. W., 1945, Résumé of facts and opinions on sedimentation in Gulf Coast region of Texas and Louisiana: Am. Assoc. Petroleum Geologists Bull., v. 29, p. 1304–1335.

Terzaghi, K., and Peck, Ralph B., 1948, Soil mechanics in engineering practice: N. Y., John Wiley and Sons, 566 p.

Trask, Parker D., 1953, Chemical studies of sediments in the Northwest Gulf of Mexico: Mass. Inst. Technology and Woods Hole Oceanographic Inst., Papers in Physical Oceanography and Meteorology, v. 12, no. 4, pt. II, p. 47–120.

Veatch, A. C., and Smith, Paul A., 1939, Atlantic submarine valleys of the United States and the Congo submarine valley: Geol. Soc. America Special Papers, no. 7, 101 p.

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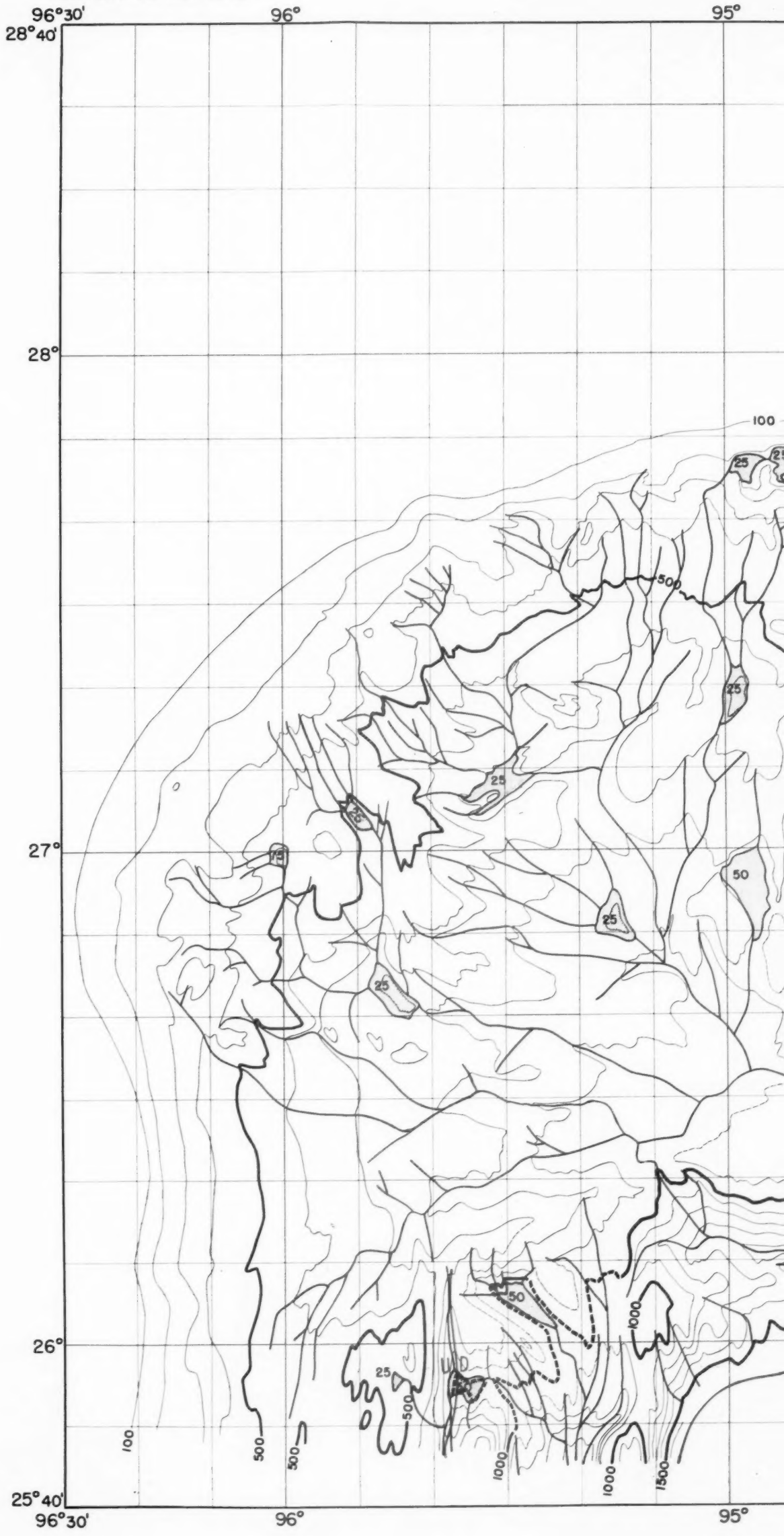
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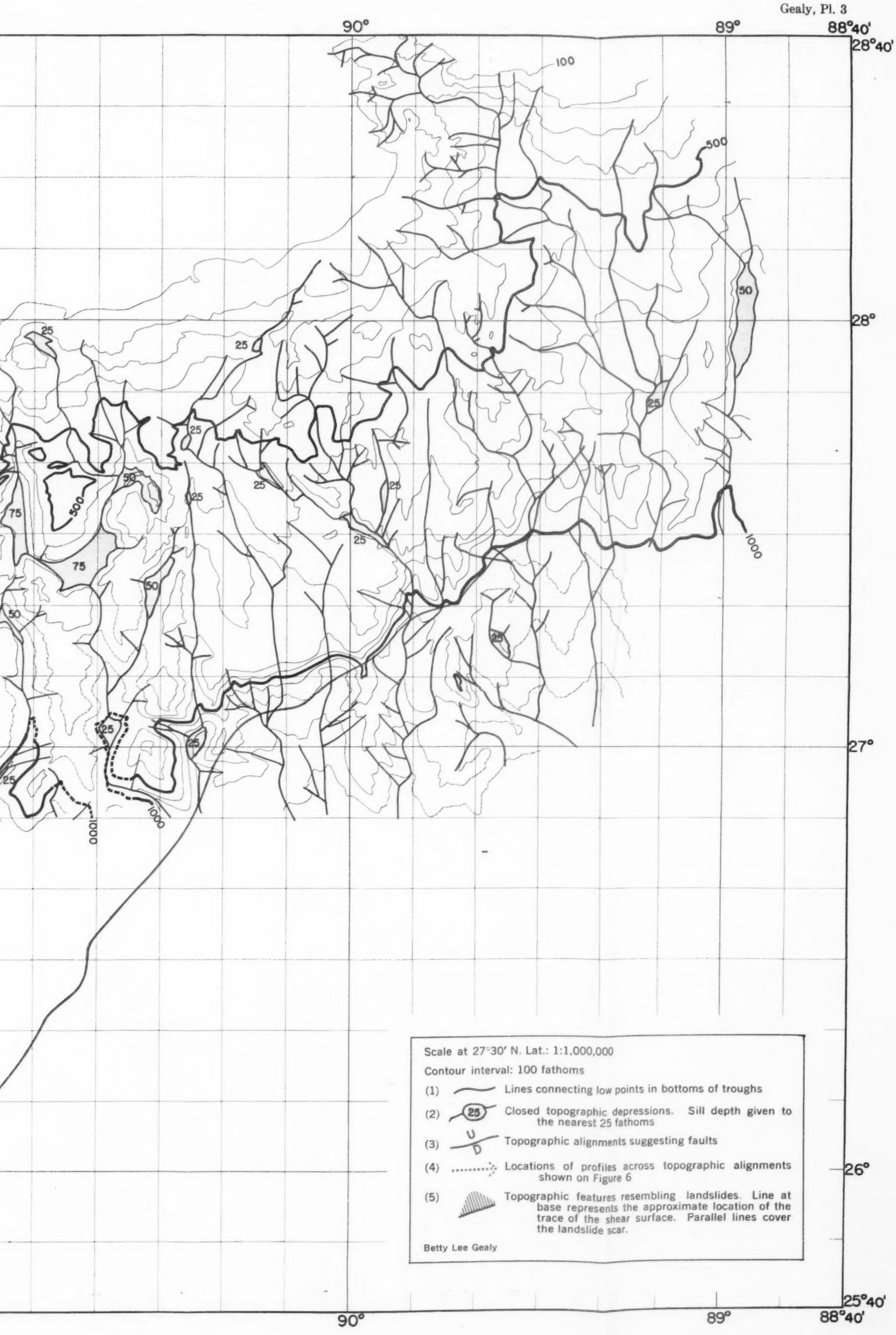
PHYSIOGRAPHIC FEATURES ON THE CONT

C FEATURES ON THE CONTINENTAL SLOPE IN THE NORTHWEST GULF OF MEXICO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Scale at 27°30' N. Lat.: 1:1,000,000
Contour interval: 100 fathoms
(1) Lines connecting low points in bottoms of troughs
(2) Closed topographic depressions. Sill depth given to the nearest 25 fathoms
(3) Topographic alignments suggesting faults
(4) Locations of profiles across topographic alignments shown on Figure 6
(5) Topographic features resembling landslides. Line at base represents the approximate location of the trace of the shear surface. Parallel lines cover the landslide scar.
Betty Lee Gealy

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SAMPLING IN BIOGEOCHEMISTRY

BY HARRY V. WARREN, ROBERT E. DELAVAULT, AND JOHN A. C. FORTESCUE

ABSTRACT

Sampling is of vital importance in biogeochemical investigations. In particular:

(1) Only organs of the same age should be compared.

(2) Preferably that organ of a tree which responds best to the element being investigated should be selected.

(3) Second-year twigs give more reproducible results than do first-year twigs.

(4) Lower and top branches may not contain the same amounts of a metal.

(5) A few specific elements are concentrated in larger amounts in particular plants: where practicable the use of these accumulator plants simplifies operations and provides better contrast between normal and anomalous samples.

(6) Anomalies are better detected when results are expressed in p.p.m. of ash rather than of dry plant.

Analyses of plants collected on the property of the Graham Bosquet Mines near Vancouver, B. C. graphically illustrate some of the above points.

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INTRODUCTION

Sampling is of vital importance in biogeochemical investigations. Just as grab, chip, channel, and bulk sampling each has its sphere of usefulness for prospectors and geologists, similarly corresponding methods have their place in biogeochemistry. Mine estimates are not made on the basis of any random set of grab samples: biogeochemical anomalies should not be sought except after systematic sampling. This paper attempts to demonstrate how sampling can make or mar a biogeochemical investigation.

Acknowledgments

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Position on Tree

It might seem reasonable to expect all parts of a tree to have a similar metal content; however, they do not. The following examples indicate that zinc appears to be present in appreciably smaller amounts in the crowns of trees than in the lower boughs.

In this and previous papers "first-year" and "second-year" growth designate the youngest and second youngest growth, respectively, at the tips of boughs: that is, these terms refer to position on a branch and not to actual age.

In Table 1 the zinc values of some second-year stems illustrate the magnitude of variations which may be expected if topmost and lower branches are sampled at random.

Sample No. 1 represents a typical negative sample, and No. 2 to No. 7 represent a line of samples taken across a narrow zone of known mineralization located near samples No. 2 and 3.

If 90(2100) p.p.m. of zinc be considered normal for this organ and species, (unpublished data—presented at Annual Meeting of Geol. Soc. America, November 1953) then the topmost twigs do not appear anomalous whereas the lower branches of samples No. 2 and 3, containing more than 50 per cent over the normal amount, are definitely anomalous.

Table 1 also illustrates the advisability of using the ash rather than the dry material content of vegetation when searching for anomalies.

Table 2 gives several analyses of other species. The samples have not been specially selected by the authors, but by other collectors at various places in Eastern Canada. The results reaffirm repeated warnings about the importance of careful and methodical sampling.

Table 1.—Balsam Fir (*Abies balsamea*) Second Year Stems; p.p.m. Zinc—Normal: 90(2100)*

Sample No.	Topmost branches	Lower branches
1	59(1700)†	70(2400)
2	55(2400)	119(3300)
3	44(1900)	77(3700)
4	77(1800)	57(1600)
5	63(1500)	83(2600)
6	45(2200)	72(2400)
7	55(1700)	57(2100)

* These and following "normals" are from unpublished data presented at the Annual Meeting of the Geological Society of America, November 1953.

† Throughout this paper all values are given in p.p.m. (parts per million) in dry plant matter with the p.p.m. ash following bracketed.

Where conditions are good almost any sample may reveal an anomalous condition, but where conditions are marginal precise work is important to obtain useful results. However carelessly the positive birch, and to a less extent the cherry, had been sampled, mineralization would have been suspected; this is clearly not the case with the spruces.

The terms "positive" and "negative" are used, as is customary in biogeochemical work, to signify the presence or absence of significant mineralization, in this case, of zinc.

Age of Sample

Many of the early workers apparently did not realize the relative magnitude of variations that may exist among the mineral contents of the variously aged leaves and stems of any single tree. Some metals are distributed more or less evenly between differently aged organs; the distribution of others varies widely.

For maximum validity of comparisons between different samples, the authors have found it advisable to take second-year stems for the majority of their samples. First year's growth

TABLE 2.—SECOND-YEAR STEMS; P.P.M. ZINC

Nature of sample (+ or −)		Topmost branches	Lower branches
Black Spruce (*Picea mariana*)	−	36 (2100)	74 (3400)
	+	49 (2600)	106 (5200)
	Normal	75 (3200)	
White Spruce (*Picea glauca*)	These three samples taken across a mineralized zone	48 (2100)	69 (2600)
		35 (2500)	85 (3400)
		46 (1400)	61 (1900)

NOTE: This suite also illustrates the advisability of recording both ash and dry plant contents of samples.

	Normal	86 (2350)	
White or Paper Birch (*Betula papyrifera*)	−	157 (4800)	307 (12,800)
	+	368 (18900)	720 (27,200)
	+	410 (19000)	770 (31,400)
	Normal	270 (8600)	
Pin Cherry (*Prunus pennsylvanica*)	These three samples taken across a mineralized zone	57 (3100)	68 (6000)
		42 (2100)	56 (2200)
		36 (1800)	26 (2800)
	Normal	52 (1300)	

TABLE 3.—COMPARISON AMONG DIFFERENTLY AGED STEMS

Species	Nature of sample (+ or −)		Copper (p.p.m.)		Zinc (p.p.m.)	
			1st yr.	2nd yr.	1st yr.	2nd yr.
Jack Pine (*Pinus Banksiana*)	−Cu	−Zn	3 (120)	5 (250)	61 (2100)	67 (3700)
	−Cu	−Zn	2 (80)	3 (180)	48 (1700)	59 (3100)
	−Cu	−Zn	3 (80)	4 (200)	34 (860)	46 (2000)
	−Cu	−Zn	2 (60)	5 (230)	81 (2300)	75 (3200)
	−Cu	−Zn	4 (150)	4 (260)	72 (2800)	70 (4100)
	−Cu	+Zn	3 (180)	6 (280)	85 (5100)	130 (7600)
	−Cu	+Zn	3 (190)	8 (290)	140 (10,000)	310 (12,000)
			Normal 5 (280)		64 (3500)	
Balsam Fir (*Abies balsamea*)	−Cu	−Zn	3 (60)	9 (220)	103 (2200)	125 (4400)
	−Cu	−Zn	5 (120)	7 (170)	59 (1300)	83 (1900)
	−Cu	−Zn	5 (150)	5 (160)	60 (1600)	80 (2600)
	+Cu	+Zn	13 (310)	14 (410)	144 (3500)	192 (5500)
	+Cu	+Zn	18 (470)	20 (680)	112 (2900)	123 (4100)
			Normal 8 (180)		90 (2100)	
White or Paper Birch (*Betula papyrifera*)	−Cu	−Zn	5 (110)	6 (230)	222 (4600)	217 (8500)
	−Cu	−Zn	6 (80)	7 (190)	168 (2300)	205 (5000)
	−Cu	−Zn	13 (200)	5 (230)	437 (6600)	212 (9900)
	+Cu	+Zn	25 (620)	36 (1800)	620 (15000)	650 (33000)
	+Cu	+Zn	22 (520)	22 (920)	610 (14000)	650 (26000)
			Normal 8 (260)		270 (8600)	
Pin Cherry (*Prunus pennsylvanica*)	−Cu	−Zn	5 (73)	6 (120)	53 (710)	67 (1300)
	−Cu	−Zn	3 (53)	3 (60)	70 (1200)	110 (2600)
	+Cu	+Zn	3 (150)	2 (180)	57 (2600)	66 (4900)
	+Cu	+Zn	12 (220)	13 (400)	287 (5400)	312 (9200)
	+Cu	+Zn	22 (340)	23 (450)	445 (6600)	600 (12,000)
			Normal 4 (100)		52 (1300)	

tends to vary in its mineral content during the season and, moreover, owing to the comparatively rapid growth of terminal leaves and stems, it is difficult to take similar samples in two separated areas during one season, or to make valid comparisons between one set taken in June and another taken in September, or two sets collected on opposite slopes of one valley or hill.

Table 4.—Nickel Content of Various Organs in p.p.m. Dry Plant

Samples taken near nickel mineralization.

Species	Twigs				Needles			
	Year				Year			
	1	2	3	4 + 5	1	2	3	Mixed
Tsuga Mertensiana	6	5	6	5				15
(Mountain Hemlock)	6	5	3	6				4
	5	5	5	5				9
Abies lasiocarpa	25	30	30	40	22	36	25	
(Mountain Balsam)	15	12	12	12	12	9	10	

Table 3 illustrates the importance of not mixing the age of the plant organs which are being sampled, at least when copper and zinc are being investigated. On the contrary, where nickel is being studied, the exact age of the organ seems of little importance (Table 4).

Collection at Different Periods of Growing Season

The authors early in their investigations realized that very young growth varies in mineral content from more mature growth. Second-year stems were therefore preferred to first-year growth, either stems or leaves. Recently agriculturists have shown that the content, at least of zinc in leaves declines steadily as the season advances (Smith and Taylor, 1952, p. 37). The same investigators showed that all elements do not behave alike. Although they were dealing with leaves of peach and apple trees, it is probably safe to infer that comparable results would be obtained with the forest species with which most biogeochemists are concerned.

Most collections of material normally are made over a short period so that this particular problem is not serious. Nevertheless, it indicates the difficulty of setting up normals and may explain why, in setting up normals, it has usually been thought necessary to have a sample differ quite widely—from 30 to 50 per cent above normal in the case of zinc—before it is considered anomalous. Where a local set of conditions is being considered, a much more specific normal can be established.

Table 5.—Metal Content of Dry Plant Material, Second-year Stems; p.p.m.

A. Manganese					
Labrador Tea (*Ledum groenlandicum*)	800	1000	1250	1500	2125
Black Spruce (*Picea mariana*)	175	300	225	300	1500
B. Zinc					
Red Alder (*Alnus rugosa*)	125	106	86	71	37
Willow (*Salix* sp.)	410	297	440	179	141
C. Copper					
Red Alder (*Alnus rugosa*)	22	19	17	15	24
Willow (*Salix* sp.)	11	10	12	11	13

Sampling of Nodes

The little knobs on branches, called nodes, conveniently mark off each period of growth. The authors recommend that the first knob nearest the tip be taken with the second-year stem because this knob may be thought of as representing the end of the previous season's growth. Tips of stems ahead of the first node constitute the latest growth.

Unfortunately, nobody, so far, seems to have made a study of the mineral content of nodes. A preliminary investigation has indicated that the amount of some elements in a node differs markedly from that in the twig on either side. Not having enough reliable analyses available,

the authors of this paper can say only that in their experience the first node usually contains less copper and a little less zinc and ash than do this node and the remainder of the second-year growth if taken together. This variation should cause no sampling problem whatsoever as long as a sampler remembers to sample systematically and uniformly.

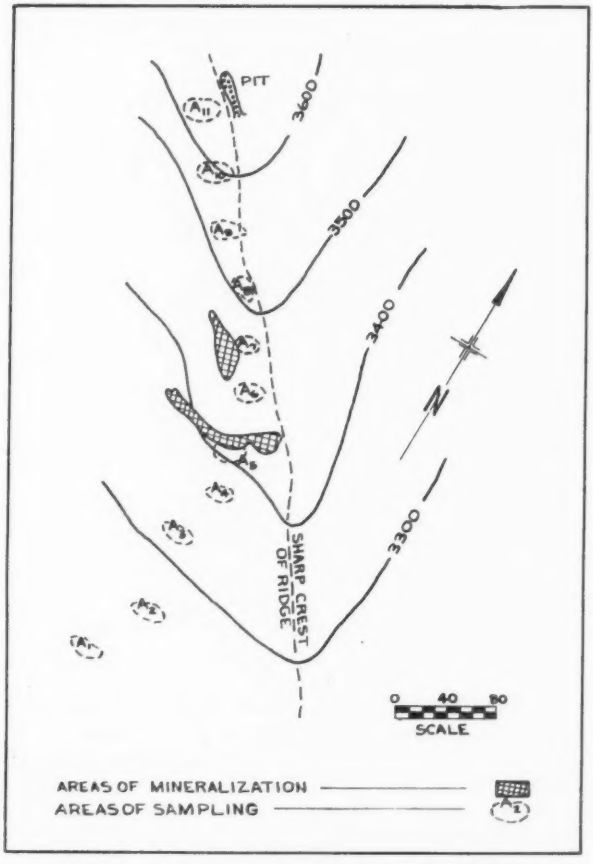

FIGURE 1.—PLAN OF GRAHAM BOSQUET ZINC SHOWINGS

SELECTING SUITABLE SPECIES

Both collecting and analyzing can be greatly simplified by choosing suitable plant species. Facts are too meager to permit listing the most suitable species of plants to be used in a search for each metal. However, the following illustrations show how, for some elements, the sample need be only from one third to one tenth of what it would be were an unsuitable species selected.

In Table 5 some examples show how, by the selection of suitable species, smaller samples are needed, and consequently collecting and analyzing costs can be cut. In this table only dry plant contents are given, but if the ash contents were added they would merely underline these results. Insofar as possible all the pairs of samples came from the same soil.

EXAMPLE OF THE IMPORTANCE OF SYSTEMATIC SAMPLING

General

During the winter of 1952–1953 Fortescue studied the biogeochemistry of a zinc prospect

near Vancouver. This prospect was under development by Graham Bousquet Gold Mines Limited and lay on a steep hillside where enough work had been done to show where mineralization occurred but where with reasonable care it was still possible to take unsalted samples. Hereafter this prospect will be termed "the Prospect". Four species of plant were sampled: Western Red Cedar (*Thuya plicata*), Western Hemlock (*Tsuga heterophylla*), Balsam Fir (*Abies amabilis*), and Blueberry (*Vaccinum ovalifolium*). Each cedar, balsam, and hemlock sample was divided into four organs for analyzing, and each of the blueberry samples into three.

The analyses provide an ideal illustration of the importance of systematic sampling.

Location

The prospect is at an elevation of about 3500 feet, 1 mile north of Hanes Creek, a tributary of Lynn Creek from which the city of Vancouver receives a portion of its water supply.

All the samples were taken from "The Upper Showings" which lie on the west side of a well-marked nose or ridge, the axis of which runs approximately north-south.

The accompanying sketches show in plan (Fig. 1) and in elevation (Fig. 2) the distribution of the mineralization and the areas from which each sample was taken. No mineralization is known between A 7 and A 11: that of the old prospect pit, shown to the east of A 11 and lying on the east side of the ridge, could not conceivably affect the sample. There are no underground workings at these upper showings.

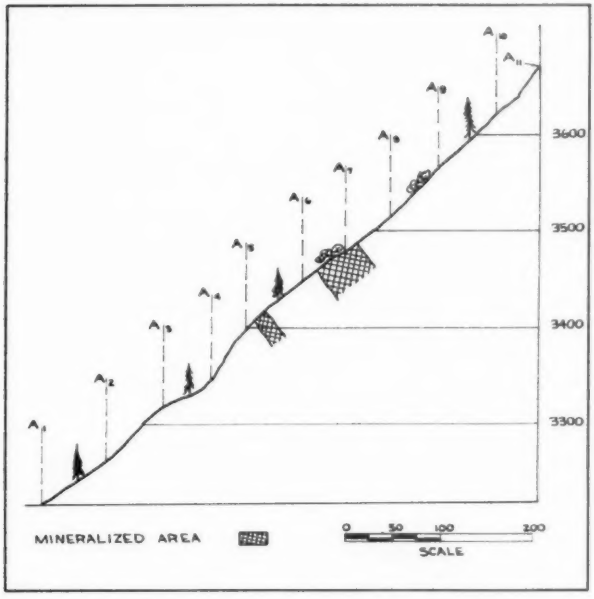

FIGURE 2.—ELEVATION OF GRAHAM BOSQUET ZINC SHOWINGS

Precipitation

No precise figures are available, but, on the basis of elevation and topographical relationship to the nearest precipitation station at Seymour Falls, it may be assumed that the annual precipitation exceeds 160 inches.

Overburden

There is seldom more than a few inches of overburden, little soil development, and bedrock outcrops in several places down the slope.

Geology and Mineralogy

Bedrock consists of fractured limestone interbedded with argillaceous bands. Mineralization consists of sphalerite with minor and erratically distributed chalcopyrite scattered along a mineralized zone.

Vegetation

The most prominent trees are large old Western Hemlock (*Tsuga heterophylla*) and Western Red Cedar (*Thuja plicata*), trees which in this area have few branches lower than 40 feet from the ground. Samples were collected from occasional smaller trees, usually less than 20 feet in height. Other samples were taken from Balsam Fir (*Abies amabilis*), some of which were small trees, and from two species of Blueberry (*Vaccinium ovalifolium* and *Vaccinium perifolium*), hereafter referred to as *Vaccinium ovalifolis*.

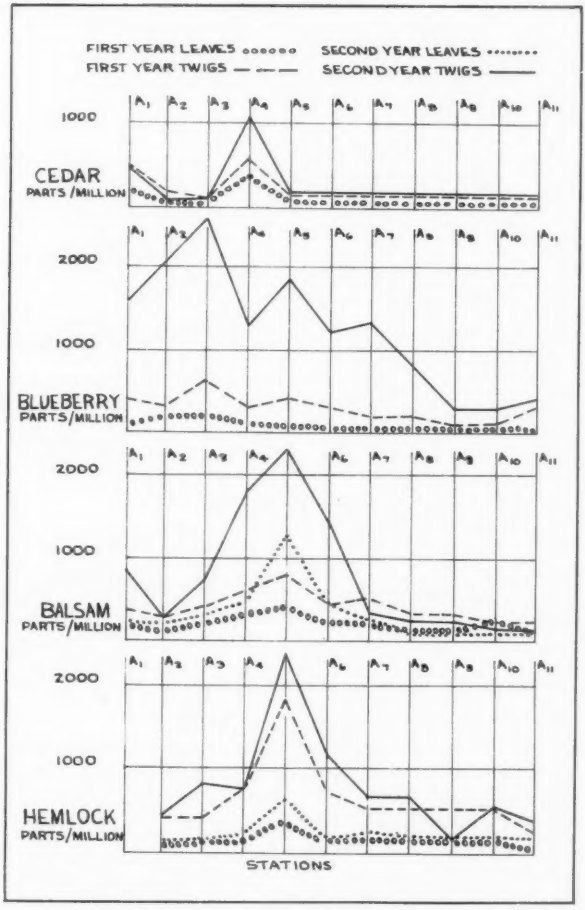

FIGURE 3.—ZINC ANALYSES OF VARIOUS ORGANS (P.P.M. ASH)

Sampling in the Field

A line was selected (Fig. 1) with 11 stations marked on it, approximately 50 feet apart. Samples were collected from trees and bushes

as near as possible to each station. In a few extreme cases a sample was taken as far as 20 feet to one side of a station. All samples were collected from trees growing on the west side of the ridge's crest.

was measured from the tip of a twig to a point near, but not including, the youngest node. Second-year growth included the youngest node and extended to a point just forward of the second node. Blueberry samples were

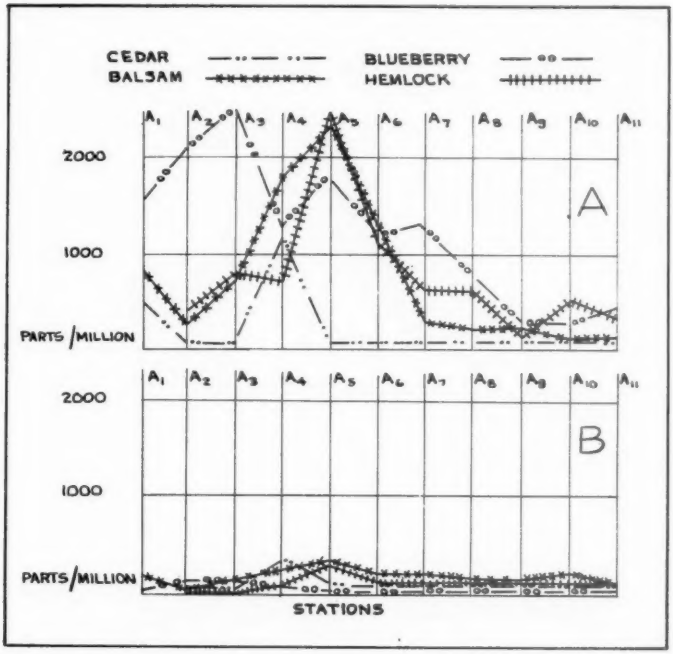

FIGURE 4.—ZINC ANALYSES OF "A" SECOND-YEAR STEMS AND "B" FIRST-YEAR LEAVES (P.P.M. ASH)

Each sample was collected in as uniform a manner as possible from branches around an individual tree. However, several blueberry bushes had to be selected to obtain an adequate sample. Sampling took place in October when growth for the year had probably ended and when the zinc content of the leaves was probably at a minimum.

Sampling in the Laboratory

As soon as the samples were received in the laboratory they were unpacked, oven dried, cut up, and separated into various organs.

The hemlock and balsam samples were divided into first- and second-year needles and first- and second-year twigs. First-year growth treated similarly except that there were only first-year leaves to separate. Because nodes are not obvious in cedar twigs, "first-year growth" was arbitrarily taken to be the green part of the stem, and "second-year growth" was taken from the light-brown portion of the stem. Probably the second-year growth actually included some older stems. Fortunately, however, the sampling was reasonably uniform.

Chemical Analyses

Analysis was carried out by procedures already described (Warren and Delavault, 1949, p. 537–540; Warren, Delavault, and Irish, 1952, p. 437–438).

Analytical Results

Graphs illustrate the results obtained by the analyses. Wherever necessary minor modifications have been made in the graphs to improve legibility. In all cases the metal values are expressed as parts per million of ash.

Figure 3 represents all the zinc analyses and it can readily be seen how in this case second-year twigs are by far the more satisfactory organ for detecting the mineralization. All first-year leaves and needles are much less satisfactory. Such might not have been the case if these samples had been taken earlier in the year. The anomaly is shifted downhill in the case of blueberry probably because of some sliding downhill of surface soil.

Figure 4 emphasizes how widely differing results can be obtained by selecting different organs, and how cedar carries much less zinc than do the other species.

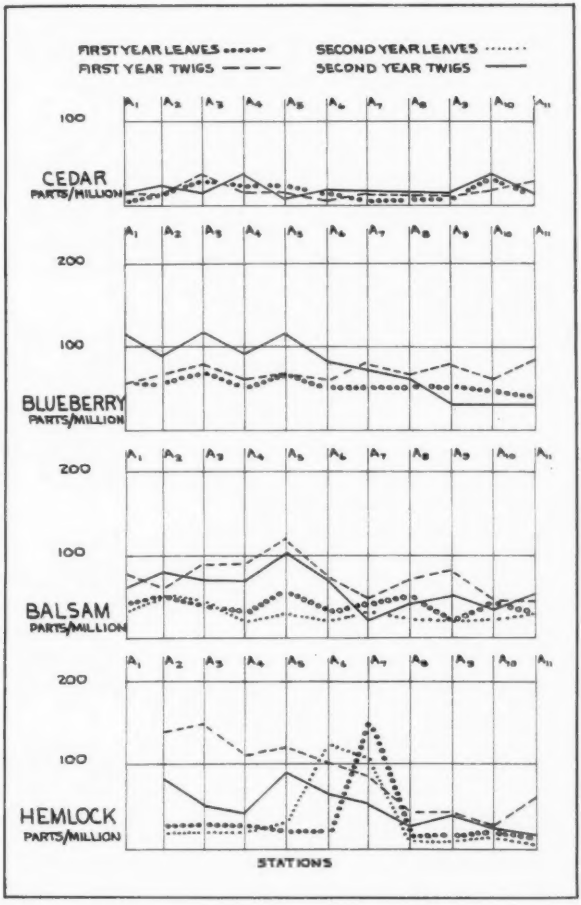

FIGURE 5.—COPPER ANALYSES OF VARIOUS ORGANS (P.P.M. ASH)

Figure 5 represents all the copper analyses and illustrates both a moderate increase in the copper content of the vegetation below the showings and the lack of any well-marked anomaly such as was shown for the zinc. This accords well with the known mineralization. Twigs appear to be slightly more sensitive

than leaves or needles to copper, but there is no marked distinction between first- and second-year growth as there was in the case of zinc. Cedar appears to carry less copper than do the other species.

In spite of hemlock's excellent response to mineralization, it is usually an unpracticable species to sample because of its slow growth and slender twigs. Consequently, cedar and hemlock should never be selected if blueberry and balsam are available and either copper or zinc anomalies are being sought.

Summary and Conclusions

The selection of suitable species and organs is important, and sampling must be uniform and systematic if biogeochemistry is to be used as a satisfactory tool in either detailed or exploratory prospecting.

Suitable biogeochemical sampling at the Graham Bosquet prospect would have clearly indicated the presence of zinc mineralization and hinted at the existence of some copper. Unspecific sampling might well have been fruitless.

References Cited

Smith, Cyril B., and Taylor, George A., 1952, Tentative optimum leaf concentrations of several elements for Elberta Peach and Stayman Apple in Pennsylvania orchards: Am. Soc. for Horticultural Sci., v. 60, p. 33–41.

Warren, H. V., and Delavault, R. E., 1949, Further studies in biogeochemistry: Geol. Soc. America Bull., v. 60, p. 531–560.

Warren, H. V., Delavault, R. E., and Irish, R. I., 1952, Biogeochemical investigations in the Pacific Northwest: Geol. Soc. America Bull., v. 63, p. 435–484.

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Short Notes

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OCCURRENCE OF *CHIROTHERIUM* IN SOUTH AMERICA

BY FRANK E. PEABODY

The discovery of the pseudosuchian footprint of the genus *Chirotherium* in Mendoza, Argentina, by Carlos Rusconi (1951) is important for our understanding of land connections between North America and South America in the early Mesozoic era. The footprint occurs in the red, sandy shales of the Higueras formation, currently considered Upper Permian. Rusconi's clear photograph of the chirothere pes leaves no doubt that his species *Chirotherium higuerensis* is nearly if not actually identical with the variety of *C. barthi* occurring in the Holbrook member of the Moenkopi formation of Arizona. The latter species is slightly but not specifically different from the type of *C. barthi* of the German Bunter. The species is among the most if not the most distinctive and most clearly recorded of known tetrapod footprints.

The pes described by Rusconi is identical with that of *C. barthi* of Arizona (*cf.* Peabody, 1948, Pl. 40) in such diagnostic details as size, claw shape, proportions of digit length, disposition and shape of digital pads, and scale pattern. The seemingly double heel mark of the pes is in reality the tip of digit V strongly recurved upon the heel and obviously showing a skid mark leading to the definitive impression of the digit from behind. In chirotheres the tip of digit V may have the position shown by Rusconi's reconstruction but, as is really the case with his specimen, is often greatly recurved. A manus impression found 50 cm from the pes (hence not belonging with the set of prints represented by the pes) is compared with the manus of several Moenkopi species clearly different from *C. barthi*. The manus as described appears more amphibian than reptilian and, if chirothere, is definitely too small to be associated with the described pes.

C. barthi of Arizona occurs near the top of the type Moenkopi, a horizon currently considered uppermost Lower or lowest Middle Triassic. The marked similarity amounting to specific identity of the Argentine footprint indicates that the Argentine stratum producing the footprint and referred to the Upper Permian is a general correlative of the Moenkopi. This practical identity of highly characteristic pseudosuchian footprints also suggests a definite land connection between North America and South America during a period when faunal relations in the Middle Triassic of East Africa and Brazil were, as pointed out by Romer (1952), close enough to suggest some sort of direct continuity between the two southern continents. Discovery of identical species of chirothere footprints in North America and South America make it necessary to explain why cynodont therapsids and rhynchocephalians, which are assumed by some to have reached South America directly from Africa, did not subsequently reach North America from the south. Romer thinks that the Moenkopi does not contain evidence of therapsids or rhynchocephalians, and that this fact minimizes the significance of the available evidence for relating South American pseudosuchians with those of North America (the latter indicated largely by their footprints). However, there are some trackways in the Moenkopi suggesting rhynchocephalians (Peabody, 1948, Pl. 29) and also some which possibly represent therapsids (Pl. 45) as well as the clear indication of a rich pseudosuchian assemblage. Lack of positive proof of the presence in North America of rhynchocephalians and cynodonts may well be a result of the vagaries of preservation. Much upland country in the province of erosion existed to

the south, east, and north of the Moenkopi province of deposition. Within the Moenkopi, skeletal material of any reptile including the patently abundant pseudosuchians is exceedingly rare; and even the extensive fauna recorded on the many mudflats represents mostly a single environment (riparian). In general support of the contention that the record of North American reptiles of Middle through Late Permian and Triassic times is markedly biased, we may note recent osteological discoveries of a large carnivorous therapsid (*Dimacrodon hottoni*) in the Middle Permian of Texas and of typical tritylodont therapsids in the latest Triassic (or early Jurassic) of Northern Arizona. These discoveries supplement the isolated and seemingly anomalous occurrence of the dicynodont *Placerias* in the Late Triassic of Arizona. Thus, evidence is accumulating that makes increasingly possible the thesis that a Holarctic distribution of early reptiles with southward dispersals could account for the composition of South American reptile faunas particularly during the Triassic. In any case, it is clear that typical chirotheres of wide distribution in the Triassic of Europe and North America were able to reach South America probably via a Central American route. Evidence of paleogeography (Woodring, 1954) strongly suggests that the concept of a Tethys sea forming a major barrier between the two continents at least during Early and Middle Triassic is untenable. An alternate proposition that chirotheres entered South America from Europe via Africa is perhaps possible but not probable in our present state of knowledge. Yet another alternate proposition that chirotheres and other pseudosuchians originated in the southern hemisphere and dispersed northward is not likely, but if true would not affect the premise that a definite land connection existed between North America and South America in Triassic time.

In summary, the occurrence of the chirothere footprint, *Chirotherium barthi*, in Argentina taken with slowly growing correlative evidence increases the possibility of explaining the dispersal of early terrestrial reptiles in terms of the Matthew-Simpson principles of mammalian dispersal.

References Cited

Peabody, F. E., 1948, Reptile and amphibian trackways from the Lower Triassic Moenkopi formation of Arizona and Utah: Univ. Calif. Pub., Bull. Dept. Geol. Sci., v. 27, p. 295-468, 40 figs., 23 pls.

Romer, A. S., 1952, [Discussion of, "The Mesozoic Tetrapods of South America" by E. H. Colbert.] Am. Mus. Nat. Hist. Bull., v. 99, no. 3, p. 250–254.

Rusconi, Carlos, 1951, Rastros de patas de reptiles Pérmicos de Mendoza: Rev. Soc. Hist. Geog. Cuyo, v. 3, no. 3, p. 1–14, 2 figs., 1 pl.

Woodring, W. P., 1954, Caribbean land and sea through the ages: Geol. Soc. America, Bull., v. 65, p. 719–732, 3 figs., 1 pl.

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